Satellite Radar Altimetry of Sea Ice

by Seymour William Clarke Laxon

A thesis submitted to the University of London for the degree of Doctor of Philosophy

Mullard Space Science Laboratory
Department of Physics and Astronomy
University College London

November 1989
Frontispiece: Antarctic sea ice extent from February 1987 to February 1988 as mapped by the Geosat radar altimeter.
The thesis concerns the analysis and interpretation of data from satellite borne radar altimeters over ice covered ocean surfaces. The applications of radar altimetry are described in detail and consider monitoring global climate change, the role that sea ice plays in the climate system, operational applications and the extension of high precision surface elevation measurements into areas of sea ice. The general nature of sea ice cover is discussed and a list of requirements for sea ice monitoring is provided and the capability of different satellite sensors to satisfy needs is examined. The operation of satellite borne altimeter over non-ocean surfaces is discussed in detail. Theories of radar backscatter over sea ice are described and are used to predict the radar altimeter response to different types of sea ice cover. Methods employed for analysis of altimeter data over sea ice are also described.

Data from the Seasat altimeter is examined on a regional and global scale and compared with sea ice climatology. Data from the Geosat altimeter is compared with co-incident imagery from the Advanced Very High Resolution Radiometer and also from airborne Synthetic Aperture Radar. Correlations are observed between the altimeter data and imagery for the ice edge position, zones within the ice cover, new ice and leads, vast floes and the fast ice boundary. An analysis of data collected by the Geosat altimeter over a period of more than two years is used to derive seasonal and inter-annual variations in the total Antarctic sea ice extent. In addition the retrieval of high accuracy elevation measurements over sea ice areas is carried out. These data are used to produce improved maps of sea surface topography over ice-covered ocean and provide evidence of the ability of the altimeter to determine sea ice freeboard directly. In addition the changing freeboard of two giant Antarctic tabular icebergs, as measured by the Geosat altimeter, is presented. As a summary the achievements are reviewed and suggestions are made towards directions for further work on present data sets and for future data from the ERS-1 satellite.
PREFACE

I would like to acknowledge the contribution of the people who have helped and guided me during my period of study towards the work described in this thesis. I am greatly indebted to my supervisor, Dr. Chris Rapley, for his constant encouragement and guidance both in matters scientific and in the wider aspects of pursuing academic research. I would also like to thank Dr. Neil McIntyre for his encouragement and help in matters glaciological, Dr. Wyn Cudlip for discussions on the operation of the altimeter and for his invaluable efforts with much of the software used at MSSL, and to Dr. Duncan Wingham, for encouraging a physical approach to remote sensing at an important early stage in my study. I would also like to thank all the past and present members of the remote sensing group at MSSL for enabling me to pursue my study in a constantly stimulating environment. I am also grateful to Len Culhane and to the department of Physics and Astronomy at UCL for allowing me to pursue my study at MSSL.

I must also acknowledge the contributions of several outside institutions towards the work presented in this thesis.

Thanks are due to the Norwegian Polar Institute for allowing me to participate in the Lance cruise, and to the other members of the cruise for making it a very memorable experience.

I would also like to express thanks to the workers at the Navy Oceans Research and Development Activity (NORDA) in Mississipi, particularly Dr. Don Johnson, Dr. Jeff Hawkins and Dr. Florence Fetterer, for allowing me to spend time at their institution and to thank Fred Abell and Needer Chase for their efforts in data processing for a very demanding visitor.

I would also like to acknowledge the contributions to this work by the Algorithm Development Facility at the Earth Observation Data Centre, Farnborough, for their efforts in providing the Polar Reference Data Set and the Geosat GDR data.

The work presented in this thesis is my own, unaided work, except where otherwise acknowledged.
TO MY PARENTS
Chapter 1 Applications of Satellite Radar Altimeter Observations Over Sea Ice

1.0 Introduction

1.1 Monitoring global climate change
   1.1.1 Components of the climate system
   1.1.2 Global climate models
   1.1.3 Climate feedback mechanisms
   1.1.4 The possibility of Global climate change
   1.1.5 Monitoring Global climate change
   1.1.6 Sea ice as an indicator of climate change

1.2 The role of sea ice in regional and global meteorology
   1.2.1 Sea ice-Ocean-Atmosphere interaction
   1.2.2 Effects of sea ice on the atmosphere
   1.2.3 Sea ice and global ocean circulation
   1.2.4 Sea ice data important in determining influences on climate

1.3 Sea ice models

1.4 Operational applications
   1.4.1 Requirements for operational sea ice information
<table>
<thead>
<tr>
<th>LIST OF CONTENTS (continued)</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.4.2 Use of satellite data for ice information and forecasting</td>
<td>35</td>
</tr>
<tr>
<td>1.4.3 Operational uses of data from the Geosat altimeter</td>
<td>36</td>
</tr>
<tr>
<td>1.4.4 Operational sea ice products from ERS-1</td>
<td>36</td>
</tr>
<tr>
<td>1.4.5 Sea ice parameters for operational applications</td>
<td>36</td>
</tr>
<tr>
<td><strong>1.5 Polar oceanography and Geophysics</strong></td>
<td></td>
</tr>
<tr>
<td>1.5.1 Measurements provided by satellite radar altimeters</td>
<td>37</td>
</tr>
<tr>
<td>1.5.2 Altimeter measurements of sea surface height</td>
<td>37</td>
</tr>
<tr>
<td>1.5.3 Altimeter measurements of ocean waves</td>
<td>38</td>
</tr>
<tr>
<td>1.5.4 Extension of altimetry into polar regions</td>
<td>40</td>
</tr>
<tr>
<td><strong>1.6 Summary</strong></td>
<td>40</td>
</tr>
</tbody>
</table>

### CHAPTER 2 SEA ICE CHARACTERISTICS AND OBSERVATIONAL TECHNIQUES

| 2.0 | Introduction | 43 |
| 2.1 | Physical characteristics of sea ice | 43 |
| 2.2 | The nature of global sea ice cover | 46 |
| 2.2.1 Differences in Arctic and Antarctic sea ice cover | 46 |
| 2.2.2 Zones observed in Arctic seasonal sea ice cover | 50 |
| 2.2.2.1 Fast ice | 50 |
| 2.2.2.2 The shear zone | 51 |
| 2.2.2.3 The marginal ice zone | 51 |
| 2.2.3 Zones observed in Antarctic sea ice cover | 53 |
| 2.3 | Data requirements for sea ice monitoring | 54 |
| 2.4 | Remote sensing of sea ice | 55 |
| 2.4.1 Visible/Infra-red observations of sea ice | 58 |
| 2.4.2 Passive microwave observations of sea ice | 60 |
| 2.4.3 Synthetic Aperture Radar observations of sea ice | 62 |
| 2.4.4 Radar altimetry over sea ice | 64 |
| 2.4.4.1 Analysis of GEOS-3 data | 64 |
## LIST OF CONTENTS (continued)

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.4.4.2 Ice freeboard measurements using satellite altimetry</td>
<td>65</td>
</tr>
<tr>
<td>2.4.4.3 Satellite radar altimeter observations of swell penetration</td>
<td>66</td>
</tr>
<tr>
<td>2.4.4.4 Physical interpretation of radar altimeter return</td>
<td>66</td>
</tr>
<tr>
<td>waveforms over sea ice</td>
<td></td>
</tr>
<tr>
<td>2.4.4.5 Radar altimeter observations during MIZEX-84</td>
<td>67</td>
</tr>
<tr>
<td>2.4.4.6 Comparisons of Radar altimeter and Synthetic Aperture Radar</td>
<td>69</td>
</tr>
<tr>
<td>observations of sea ice</td>
<td></td>
</tr>
<tr>
<td>2.4.4.7 Comparisons of Seasat altimeter data with ice charts</td>
<td>70</td>
</tr>
<tr>
<td>2.4.4.8 Comparisons of airborne altimetry with visible imagery</td>
<td>70</td>
</tr>
<tr>
<td>2.4.4.9 Operational use of Geosat radar altimeter data over sea ice</td>
<td>71</td>
</tr>
<tr>
<td>2.4.4.10 Previous work on altimetry over sea ice - discussion</td>
<td>72</td>
</tr>
<tr>
<td>2.5 Satellite instrument capability for providing sea ice parameters</td>
<td>73</td>
</tr>
<tr>
<td>2.6 Summary</td>
<td>74</td>
</tr>
</tbody>
</table>

### CHAPTER 3 THE SEASAT ALTIMETER

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.0 Introduction</td>
<td>75</td>
</tr>
<tr>
<td>3.1 Principles of radar altimetry</td>
<td>75</td>
</tr>
<tr>
<td>3.1.1 Return pulse profile</td>
<td>75</td>
</tr>
<tr>
<td>3.1.2 Modes of operation</td>
<td>76</td>
</tr>
<tr>
<td>3.1.3 Waveform averaging</td>
<td>76</td>
</tr>
<tr>
<td>3.2 The Seasat altimeter system</td>
<td>77</td>
</tr>
<tr>
<td>3.2.1 Full deramp processing</td>
<td>77</td>
</tr>
<tr>
<td>3.2.2 Seasat return waveform sampling</td>
<td>78</td>
</tr>
<tr>
<td>3.3 Pulse limited altimeter geometry</td>
<td>79</td>
</tr>
<tr>
<td>3.3.1 Range ring geometry</td>
<td>79</td>
</tr>
<tr>
<td>3.3.2 Altimeter footprints</td>
<td>80</td>
</tr>
<tr>
<td>3.3.2.1 Pulse limited footprint</td>
<td>80</td>
</tr>
<tr>
<td>3.3.2.2 Range window footprint</td>
<td>80</td>
</tr>
<tr>
<td>3.3.2.3 Beam limited footprint</td>
<td>80</td>
</tr>
</tbody>
</table>
### List of Contents (continued)

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.4</td>
<td>Seasat altimeter on-board processing</td>
<td>81</td>
</tr>
<tr>
<td>3.4.1</td>
<td>Seasat control loops</td>
<td>81</td>
</tr>
<tr>
<td>3.4.2</td>
<td>AGC loop processing</td>
<td>81</td>
</tr>
<tr>
<td>3.4.3</td>
<td>Height loop processing</td>
<td>82</td>
</tr>
<tr>
<td>3.5</td>
<td>Processing of non-ocean like waveforms</td>
<td>85</td>
</tr>
<tr>
<td>3.5.1</td>
<td>Peaked return AGC loop processing</td>
<td>85</td>
</tr>
<tr>
<td>3.5.2</td>
<td>Peaked return height loop processing</td>
<td>86</td>
</tr>
<tr>
<td>3.5.3</td>
<td>Telemetry summing</td>
<td>86</td>
</tr>
<tr>
<td>3.6</td>
<td>Geometric effects</td>
<td>87</td>
</tr>
<tr>
<td>3.6.1</td>
<td>Antenna mispointing</td>
<td>87</td>
</tr>
<tr>
<td>3.6.2</td>
<td>Topographic variations</td>
<td>88</td>
</tr>
<tr>
<td>3.6.3</td>
<td>Backscatter variations</td>
<td>89</td>
</tr>
<tr>
<td>3.7</td>
<td>The ERS-1 altimeter</td>
<td>89</td>
</tr>
<tr>
<td>3.8</td>
<td>Summary</td>
<td>90</td>
</tr>
</tbody>
</table>

### Chapter 4 Normal Incidence Radar Backscatter Over Sea Ice

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.0</td>
<td>Introduction</td>
<td>92</td>
</tr>
<tr>
<td>4.1</td>
<td>Measurement of radar backscatter by a satellite altimeter</td>
<td>93</td>
</tr>
<tr>
<td>4.2</td>
<td>Dielectric properties of sea ice</td>
<td>94</td>
</tr>
<tr>
<td>4.3</td>
<td>Reflection from a perfectly flat surface</td>
<td>96</td>
</tr>
<tr>
<td>4.4</td>
<td>Reflection from a collection of facets</td>
<td>98</td>
</tr>
<tr>
<td>4.4.1</td>
<td>Diffraction theory</td>
<td>98</td>
</tr>
<tr>
<td>4.4.2</td>
<td>Phase coherency between different facets</td>
<td>100</td>
</tr>
<tr>
<td>4.4.3</td>
<td>A facet model for normal incidence backscatter from first-year sea ice</td>
<td>101</td>
</tr>
<tr>
<td>4.5</td>
<td>Rough surface scattering</td>
<td>103</td>
</tr>
<tr>
<td>4.5.1</td>
<td>The Rayleigh criterion</td>
<td>103</td>
</tr>
<tr>
<td>4.5.2</td>
<td>Small perturbation theory</td>
<td>105</td>
</tr>
<tr>
<td>4.5.3</td>
<td>Physical optics approximation</td>
<td>106</td>
</tr>
<tr>
<td>4.5.4</td>
<td>Two scale models</td>
<td>110</td>
</tr>
<tr>
<td>Section</td>
<td>Title</td>
<td>Page</td>
</tr>
<tr>
<td>---------</td>
<td>------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>4.6</td>
<td>Empirical models and observations of radar backscatter from sea ice</td>
<td>111</td>
</tr>
<tr>
<td>4.7</td>
<td>Surface roughness in sea ice areas</td>
<td>113</td>
</tr>
<tr>
<td>4.7.1</td>
<td>Surface roughness of the ocean in sea ice areas</td>
<td>114</td>
</tr>
<tr>
<td>4.7.1.1</td>
<td>Swell attenuation by sea ice</td>
<td>115</td>
</tr>
<tr>
<td>4.7.1.2</td>
<td>Wave generation by wind in the ice pack</td>
<td>115</td>
</tr>
<tr>
<td>4.7.2</td>
<td>Surface roughness of grease ice</td>
<td>117</td>
</tr>
<tr>
<td>4.7.3</td>
<td>Surface roughness of newly formed ice</td>
<td>117</td>
</tr>
<tr>
<td>4.7.4</td>
<td>Surface roughness of ice floes</td>
<td>118</td>
</tr>
<tr>
<td>4.7.5</td>
<td>Surface roughness of meltponds</td>
<td>118</td>
</tr>
<tr>
<td>4.8</td>
<td>Discussion</td>
<td>119</td>
</tr>
<tr>
<td>4.9</td>
<td>Conclusions</td>
<td>119</td>
</tr>
</tbody>
</table>

**CHAPTER 5 DATA PROCESSING**

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>5.0</td>
<td>Introduction</td>
<td>121</td>
</tr>
<tr>
<td>5.1</td>
<td>Altimeter data formats</td>
<td>121</td>
</tr>
<tr>
<td>5.1.1</td>
<td>Seasat data formats</td>
<td>121</td>
</tr>
<tr>
<td>5.1.2</td>
<td>Geosat data formats</td>
<td>121</td>
</tr>
<tr>
<td>5.2</td>
<td>Processing of altimeter data over sea ice</td>
<td>123</td>
</tr>
<tr>
<td>5.3</td>
<td>Waveform quality control</td>
<td>124</td>
</tr>
<tr>
<td>5.4</td>
<td>Waveform sampler gain</td>
<td>126</td>
</tr>
<tr>
<td>5.5</td>
<td>Waveform retracking</td>
<td>127</td>
</tr>
<tr>
<td>5.6</td>
<td>Retrieval of backscatter co-efficient from Seasat data</td>
<td>128</td>
</tr>
<tr>
<td>5.7</td>
<td>Sea ice waveform parameterisation</td>
<td>129</td>
</tr>
<tr>
<td>5.7.1</td>
<td>Leading edge parameters</td>
<td>130</td>
</tr>
<tr>
<td>5.7.2</td>
<td>Peak backscatter value</td>
<td>132</td>
</tr>
<tr>
<td>5.7.3</td>
<td>Trailing edge parameters</td>
<td>132</td>
</tr>
<tr>
<td>5.8</td>
<td>Summary</td>
<td>134</td>
</tr>
</tbody>
</table>
# LIST OF CONTENTS (continued)

## CHAPTER 6  ANALYSIS OF SEASAT DATA

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.0</td>
<td>Introduction</td>
<td>135</td>
</tr>
<tr>
<td>6.1</td>
<td>The Seasat mission</td>
<td>135</td>
</tr>
<tr>
<td>6.2</td>
<td>Altimeter parameters over the East Greenland Sea</td>
<td>135</td>
</tr>
<tr>
<td>6.3</td>
<td>Temporal variations in altimeter data over sea ice</td>
<td>140</td>
</tr>
<tr>
<td>6.4</td>
<td>Distribution of waveform parameter values from the Polar Reference data set</td>
<td>144</td>
</tr>
<tr>
<td>6.4.1</td>
<td>Distribution of altimeter parameters in the Antarctic</td>
<td>144</td>
</tr>
<tr>
<td>6.4.2</td>
<td>Distribution of altimeter parameters in the Arctic</td>
<td>148</td>
</tr>
<tr>
<td>6.5</td>
<td>Sea ice global peak backscatter statistics</td>
<td>148</td>
</tr>
<tr>
<td>6.6</td>
<td>Conclusions</td>
<td>149</td>
</tr>
</tbody>
</table>

## CHAPTER 7  GEOSAT ALTIMETRY IN COMPARISON WITH IMAGERY

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.0</td>
<td>Introduction</td>
<td>151</td>
</tr>
<tr>
<td>7.1</td>
<td>AVHRR images used for comparison</td>
<td>151</td>
</tr>
<tr>
<td>7.2</td>
<td>Observations of compact and diffuse ice edges in altimeter and AVHRR data</td>
<td>153</td>
</tr>
<tr>
<td>7.3</td>
<td>Observations of new and compact ice in altimeter and AVHRR data</td>
<td>158</td>
</tr>
<tr>
<td>7.3.1</td>
<td>Kara Sea image</td>
<td>158</td>
</tr>
<tr>
<td>7.3.2</td>
<td>Chukchi Sea images</td>
<td>161</td>
</tr>
<tr>
<td>7.4</td>
<td>Observations of vast floes in altimeter and AVHRR data</td>
<td>161</td>
</tr>
<tr>
<td>7.5</td>
<td>Comparison of Geosat altimetry and airborne SAR data collected during MIZEX-87</td>
<td>166</td>
</tr>
<tr>
<td>7.5.1</td>
<td>Airborne SAR observations of sea ice off East Greenland</td>
<td>167</td>
</tr>
<tr>
<td>7.5.2</td>
<td>Comparison of Geosat altimeter and airborne SAR observations of sea ice off East Greenland</td>
<td>169</td>
</tr>
<tr>
<td>7.5.3</td>
<td>Discussion of SAR/Altimeter comparison</td>
<td>170</td>
</tr>
<tr>
<td>7.6</td>
<td>Conclusions</td>
<td>170</td>
</tr>
</tbody>
</table>
# LIST OF CONTENTS (continued)

## CHAPTER 8  SEASONAL AND INTERANNUAL VARIATION IN ANTARCTIC SEA ICE EXTENT MAPPED BY GEOSATALTImETRY

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>8.0</td>
<td>Introduction</td>
<td>172</td>
</tr>
<tr>
<td>8.1</td>
<td>Geosat ocean GDR global coverage</td>
<td>172</td>
</tr>
<tr>
<td>8.2</td>
<td>Data averaging prior to classification</td>
<td>174</td>
</tr>
<tr>
<td>8.3</td>
<td>Selection of suitable thresholds for sea ice detection</td>
<td>174</td>
</tr>
<tr>
<td>8.4</td>
<td>Classification of data bins</td>
<td>177</td>
</tr>
<tr>
<td>8.5</td>
<td>Comparison of total Antarctic sea ice extent measured by satellite altimetry and passive microwave observations</td>
<td>181</td>
</tr>
<tr>
<td>8.6</td>
<td>Conclusions</td>
<td>183</td>
</tr>
</tbody>
</table>

## CHAPTER 9  SATELLITE ALTIMETER ELEVATION MEASUREMENTS OVER SEA ICE

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>9.0</td>
<td>Introduction</td>
<td>184</td>
</tr>
<tr>
<td>9.1</td>
<td>Surface elevation measurements from satellite altimetry</td>
<td>184</td>
</tr>
<tr>
<td>9.2</td>
<td>Seasat elevation profiles obtained during the 3 day repeat cycle</td>
<td>186</td>
</tr>
<tr>
<td></td>
<td>9.2.1 Assessment of altimeter retracking over sea ice</td>
<td>186</td>
</tr>
<tr>
<td></td>
<td>9.2.2 Elevation profiles in comparison to peak backscatter</td>
<td>190</td>
</tr>
<tr>
<td>9.3</td>
<td>Mean surface elevation over sea ice in the Southern Ocean</td>
<td>193</td>
</tr>
<tr>
<td></td>
<td>9.3.1 Raw and retracked elevation noise</td>
<td>193</td>
</tr>
<tr>
<td></td>
<td>9.3.2 Southern ocean mean sea surface elevation</td>
<td>195</td>
</tr>
<tr>
<td></td>
<td>9.3.3 Elevation profiles off the Oates coast</td>
<td>197</td>
</tr>
<tr>
<td>9.4</td>
<td>Iceberg freeboard measurements</td>
<td>200</td>
</tr>
<tr>
<td></td>
<td>9.4.1 Iceberg identification</td>
<td>200</td>
</tr>
<tr>
<td></td>
<td>9.4.2 Iceberg freeboard determination</td>
<td>202</td>
</tr>
<tr>
<td></td>
<td>9.4.3 Changes in the mean freeboard of giant tabular</td>
<td>202</td>
</tr>
<tr>
<td></td>
<td>icebergs determined by the Geosat altimeter</td>
<td>202</td>
</tr>
<tr>
<td>9.5</td>
<td>Conclusions</td>
<td>209</td>
</tr>
</tbody>
</table>
## LIST OF CONTENTS (continued)

### CHAPTER 10 CONCLUSIONS

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>10.0</td>
<td>Introduction</td>
<td>211</td>
</tr>
<tr>
<td>10.1</td>
<td>Assessment of achievements</td>
<td>211</td>
</tr>
<tr>
<td>10.2</td>
<td>Directions for future work</td>
<td>216</td>
</tr>
<tr>
<td>10.2.1</td>
<td>Models for altimeter returns over sea ice</td>
<td>216</td>
</tr>
<tr>
<td>10.2.2</td>
<td>Altimeter/AVHRR comparison</td>
<td>216</td>
</tr>
<tr>
<td>10.2.3</td>
<td>Altimeter mapping of sea ice extent</td>
<td>217</td>
</tr>
<tr>
<td>10.2.4</td>
<td>Altimeter measurements of surface elevation in sea ice areas</td>
<td>217</td>
</tr>
<tr>
<td>10.3</td>
<td>The promise of ERS-1</td>
<td>217</td>
</tr>
</tbody>
</table>

### REFERENCES

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
</table>

### APPENDIX A OBSERVATIONS ON THE LANCE CRUISE 1986

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>A.0</td>
<td>Introduction</td>
<td>229</td>
</tr>
<tr>
<td>A.1</td>
<td>Lance cruise general observations</td>
<td>231</td>
</tr>
<tr>
<td>A.1.1</td>
<td>Water roughness</td>
<td>231</td>
</tr>
<tr>
<td>A.1.2</td>
<td>Sea ice cover</td>
<td>231</td>
</tr>
<tr>
<td>A.1.3</td>
<td>Ice floes</td>
<td>234</td>
</tr>
<tr>
<td>A.1.4</td>
<td>Meltponds</td>
<td>235</td>
</tr>
<tr>
<td>A.2</td>
<td>Ice floe surface roughness measurements</td>
<td>237</td>
</tr>
<tr>
<td>A.3</td>
<td>Meltpond observations</td>
<td>239</td>
</tr>
<tr>
<td>A.4</td>
<td>Conclusions</td>
<td>241</td>
</tr>
</tbody>
</table>

### LIST OF ICE TERMS

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
</table>

### LIST OF ALTIMETER DERIVED PARAMETERS

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
</table>

### LIST OF SYMBOLS

<table>
<thead>
<tr>
<th>Section</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
</table>
## LIST OF FIGURES

### CHAPTER 1

<table>
<thead>
<tr>
<th>Figure Number</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>Components of the climate system.</td>
<td>22</td>
</tr>
<tr>
<td>1.2</td>
<td>Predicted rise in global and polar temp due to CO\textsubscript{2} warming.</td>
<td>25</td>
</tr>
<tr>
<td>1.3</td>
<td>Variations in interannual sea ice extent from satellite passive microwave</td>
<td>27</td>
</tr>
<tr>
<td></td>
<td>observations.</td>
<td></td>
</tr>
<tr>
<td>1.4</td>
<td>Changes in mean ice thickness with CO\textsubscript{2} doubling.</td>
<td>27</td>
</tr>
<tr>
<td>1.5</td>
<td>Sea ice-ocean-atmosphere interaction at the ice edge.</td>
<td>29</td>
</tr>
<tr>
<td>1.6</td>
<td>Southern ocean circulation.</td>
<td>31</td>
</tr>
<tr>
<td>1.7</td>
<td>Processing controlling sea ice thickness.</td>
<td>32</td>
</tr>
<tr>
<td>1.8</td>
<td>Mean sea surface topography derived from the Seasat radar altimeter.</td>
<td>39</td>
</tr>
</tbody>
</table>

### CHAPTER 2

<table>
<thead>
<tr>
<th>Figure Number</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1</td>
<td>The growth of pancake and new ice.</td>
<td>44</td>
</tr>
<tr>
<td>2.2</td>
<td>Stages in the evolution of sea ice.</td>
<td>44</td>
</tr>
<tr>
<td>2.3</td>
<td>Arctic sea ice keel draught distribution.</td>
<td>45</td>
</tr>
<tr>
<td>2.4</td>
<td>Arctic minimum and maximum sea ice extent.</td>
<td>47</td>
</tr>
<tr>
<td>2.5</td>
<td>Antarctic minimum and maximum sea ice extent.</td>
<td>48</td>
</tr>
<tr>
<td>2.6</td>
<td>Arctic and Antarctic currents.</td>
<td>49</td>
</tr>
<tr>
<td>2.7</td>
<td>Sea ice distribution in the Marginal Ice Zone.</td>
<td>52</td>
</tr>
<tr>
<td>2.8</td>
<td>Ice freeboard determination using GEOS-3 altimeter data.</td>
<td>65</td>
</tr>
<tr>
<td>2.9</td>
<td>Ice zones observed in Seasat data.</td>
<td>69</td>
</tr>
</tbody>
</table>

### CHAPTER 3

<table>
<thead>
<tr>
<th>Figure Number</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1</td>
<td>Principle of pulse compression</td>
<td>77</td>
</tr>
<tr>
<td>3.2</td>
<td>Seasat return waveform gates</td>
<td>78</td>
</tr>
<tr>
<td>3.3</td>
<td>Range ring geometry</td>
<td>79</td>
</tr>
<tr>
<td>3.4</td>
<td>Ocean waveform processing</td>
<td>83</td>
</tr>
<tr>
<td>3.5</td>
<td>Peaked waveform processing</td>
<td>86</td>
</tr>
</tbody>
</table>
### LIST OF FIGURES (continued)

#### CHAPTER 4

<table>
<thead>
<tr>
<th>Figure</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.1</td>
<td>Fresnel zones for a radar altimeter</td>
<td>97</td>
</tr>
<tr>
<td>4.2</td>
<td>Electrical field at a point from an extended source</td>
<td>98</td>
</tr>
<tr>
<td>4.3</td>
<td>Waveform parameter values for the Brown facet model</td>
<td>103</td>
</tr>
<tr>
<td>4.4</td>
<td>The Rayleigh criterion</td>
<td>104</td>
</tr>
<tr>
<td>4.5</td>
<td>Peak backscatter from a slightly rough surface.</td>
<td>105</td>
</tr>
<tr>
<td>4.6</td>
<td>Waveform parameter values predicted by Specular point theory</td>
<td>107</td>
</tr>
<tr>
<td></td>
<td>(Stationary phase approximation; Gaussian height PDF).</td>
<td></td>
</tr>
<tr>
<td>4.7</td>
<td>Waveform parameter values predicted by Specular point theory</td>
<td>108</td>
</tr>
<tr>
<td></td>
<td>(Stationary phase approximation; Exponential height PDF).</td>
<td></td>
</tr>
<tr>
<td>4.8</td>
<td>Peak backscatter predicted by the Specular point theory</td>
<td>109</td>
</tr>
<tr>
<td></td>
<td>(Scalar approximation).</td>
<td></td>
</tr>
<tr>
<td>4.9</td>
<td>Measurements of radar backscatter from sea ice off Point Barrow.</td>
<td>112</td>
</tr>
<tr>
<td>4.10</td>
<td>Roughness components of sea ice covered ocean</td>
<td>113</td>
</tr>
<tr>
<td>4.11</td>
<td>Swell attenuation by sea ice</td>
<td>115</td>
</tr>
<tr>
<td>4.12</td>
<td>Wind regeneration of small waves</td>
<td>116</td>
</tr>
</tbody>
</table>

#### CHAPTER 5

<table>
<thead>
<tr>
<th>Figure</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>5.1</td>
<td>The 'height glitch' phenomenon</td>
<td>125</td>
</tr>
<tr>
<td>5.2</td>
<td>Waveform sampler correction</td>
<td>127</td>
</tr>
<tr>
<td>5.3</td>
<td>Threshold retracking technique</td>
<td>128</td>
</tr>
<tr>
<td>5.4</td>
<td>Calculation of leading edge parameters</td>
<td>130</td>
</tr>
<tr>
<td>5.5</td>
<td>Calculation of broad sea ice gates</td>
<td>133</td>
</tr>
</tbody>
</table>

#### CHAPTER 6

<table>
<thead>
<tr>
<th>Figure</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.1</td>
<td>Seasat waveform parameters over the East Greenland Sea.</td>
<td>136</td>
</tr>
<tr>
<td>6.2</td>
<td>Plot of reflection coefficient from altimeter observations of the Greenland Sea.</td>
<td>138</td>
</tr>
<tr>
<td>6.3</td>
<td>Seasat altimeter track overlayed on visible imagery.</td>
<td>139</td>
</tr>
<tr>
<td>LIST OF FIGURES (continued)</td>
<td>Page</td>
<td></td>
</tr>
<tr>
<td>-----------------------------</td>
<td>------</td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER 6 (continued)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6.4 Location of three day repeat track selected for analysis.</td>
<td>140</td>
<td></td>
</tr>
<tr>
<td>6.5 Temporal variations in altimeter backscatter.</td>
<td>142</td>
<td></td>
</tr>
<tr>
<td>6.6 Temporal variations in waveform profiles.</td>
<td>143</td>
<td></td>
</tr>
<tr>
<td>6.7 Seasat waveform parameter distributions in the Southern Ocean.</td>
<td>145</td>
<td></td>
</tr>
<tr>
<td>6.8 Seasat waveform parameter distributions in the Arctic Ocean.</td>
<td>147</td>
<td></td>
</tr>
<tr>
<td>6.9 Histogram of peak backscatter values.</td>
<td>149</td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER 7</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>7.1 Altimeter and AVHRR profiles over a diffuse ice edge (Greenland Sea)</td>
<td>157</td>
<td></td>
</tr>
<tr>
<td>7.2 Altimeter and AVHRR profiles over a new ice and leads (Kara Sea)</td>
<td>159</td>
<td></td>
</tr>
<tr>
<td>7.3 Altimeter and AVHRR profiles over vast ice floes (Greenland Sea)</td>
<td>164</td>
<td></td>
</tr>
<tr>
<td>7.4 Mean waveform profiles over vast ice floes</td>
<td>165</td>
<td></td>
</tr>
<tr>
<td>7.5 Geosat altimeter data in comparison with airborne SAR</td>
<td>168</td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER 8</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>8.1 Global Geosat Geophysical Data Record coverage</td>
<td>173</td>
<td></td>
</tr>
<tr>
<td>8.2 Geosat GDR coverage of the Southern Ocean</td>
<td>173</td>
<td></td>
</tr>
<tr>
<td>8.3 Plots of SDH, AGC and SWH</td>
<td>175</td>
<td></td>
</tr>
<tr>
<td>8.4 Histogram of SDH, AGC, SWH values</td>
<td>176</td>
<td></td>
</tr>
<tr>
<td>8.5 Classification algorithm flowchart</td>
<td>178</td>
<td></td>
</tr>
<tr>
<td>8.6 Plots of ice classification</td>
<td>180</td>
<td></td>
</tr>
<tr>
<td>8.7 Comparison total Antarctic ice extent derived from altimeter with SMMR</td>
<td>182</td>
<td></td>
</tr>
<tr>
<td><strong>CHAPTER 9</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>9.1 Three day repeat cycle elevation residuals (inward pass).</td>
<td>188-9</td>
<td></td>
</tr>
<tr>
<td>9.2 Three day repeat cycle elevation residuals (outward pass).</td>
<td>191-2</td>
<td></td>
</tr>
<tr>
<td>9.3 Standard deviation of raw and retracked elevation values in the Antarctic.</td>
<td>194</td>
<td></td>
</tr>
</tbody>
</table>
LIST OF FIGURES (continued)

CHAPTER 9 (continued)

9.4 Standard deviation of raw elevation values in the Arctic 194
9.5 Published Southern ocean bathymetry compared with mean sea surface topography derived from Seasat altimetry. 196
9.6 Altimeter elevation profiles near the Oates coast 198
9.7 Waveform parameter profiles over elevation changes off the Oates coast 199
9.8 The Larsen icebergs observed in satellite imagery 201
9.9 Identification of the Larsen A giant tabular iceberg 201
9.10 Method used to determine iceberg freeboard from collinear altimeter tracks 203
9.11 Iceberg Larsen A freeboard profiles 204
9.12 Iceberg Larsen B freeboard profiles 205
9.13 Observed changes in the freeboard of two giant tabular icebergs 206
9.14 Potential temperature maps of the Weddell Sea 208

APPENDIX A

A.1 Track of the Lance Cruise 230
A.2 Multi-year ice floe surface roughness 238
A.3 First year ice surface roughness 238
A.5 Histogram of meltpond radius 240
A.6 Histogram of meltpond elevations 240
## LIST OF PLATES

### CHAPTER 7

<table>
<thead>
<tr>
<th>Plate</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.1</td>
<td>Geosat Ice index overlayed on AVHRR : Greenland Sea 10th May 1987.</td>
<td>154</td>
</tr>
<tr>
<td>7.2</td>
<td>Geosat waveform parameters overlayed on AVHRR : Greenland Sea 10th May 1987.</td>
<td>154</td>
</tr>
<tr>
<td>7.3</td>
<td>Geosat Ice index overlayed on AVHRR : Greenland Sea 18th March 1987.</td>
<td>155</td>
</tr>
<tr>
<td>7.4</td>
<td>Geosat waveform parameters overlayed on AVHRR : Greenland Sea 18th March 1987.</td>
<td>155</td>
</tr>
<tr>
<td>7.5</td>
<td>Geosat waveform parameters overlayed on AVHRR : Greenland Sea 14th December 1987.</td>
<td>157</td>
</tr>
<tr>
<td>7.6</td>
<td>Geosat AGC and VATT values overlayed on AVHRR : Kara Sea 21st February 1986.</td>
<td>159</td>
</tr>
<tr>
<td>7.7</td>
<td>Geosat waveform parameters overlayed on AVHRR : Chukchi Sea 10th February 1988.</td>
<td>160</td>
</tr>
<tr>
<td>7.8</td>
<td>Geosat waveform parameters overlayed on AVHRR : Chukchi Sea 14th March 1988.</td>
<td>160</td>
</tr>
<tr>
<td>7.9</td>
<td>Geosat SIGPK and SIGTD parameters overlayed on AVHRR : Greenland Sea 18th March 1987.</td>
<td>163</td>
</tr>
<tr>
<td>7.10</td>
<td>Geosat LEWID and LEDIF parameters overlayed on AVHRR : Greenland Sea 18th March 1987.</td>
<td>163</td>
</tr>
</tbody>
</table>

### APPENDIX B

<table>
<thead>
<tr>
<th>Section</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>A.1</td>
<td>'Glassy' ocean surface caused by surface meltwater</td>
<td>232</td>
</tr>
<tr>
<td>A.2</td>
<td>Attenuation of ocean capillarity waves by ice floes</td>
<td>232</td>
</tr>
<tr>
<td>A.3</td>
<td>Area of continuous rotten sea ice cover</td>
<td>233</td>
</tr>
<tr>
<td>A.4</td>
<td>Smooth ocean surface within the ice pack</td>
<td>233</td>
</tr>
<tr>
<td>A.5</td>
<td>Surface roughness of a first year ice floe</td>
<td>235</td>
</tr>
<tr>
<td>A.6</td>
<td>Meltpond and ocean roughness</td>
<td>235</td>
</tr>
<tr>
<td>A.7</td>
<td>Frozen meltponds on a second-year ice floe</td>
<td>236</td>
</tr>
<tr>
<td>A.8</td>
<td>Surface roughness of a frozen meltpond</td>
<td>236</td>
</tr>
</tbody>
</table>
# LIST OF TABLES

<table>
<thead>
<tr>
<th>CHAPTER 1</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1 Basic monitoring requirements sea ice model validation</td>
<td>34</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CHAPTER 2</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1 Structure of the marginal ice zone of the Bering Sea</td>
<td>53</td>
</tr>
<tr>
<td>2.2 Zones within the Antarctic sea ice cover</td>
<td>54</td>
</tr>
<tr>
<td>2.3 Sea ice data requirements as defined by the ICEX working group</td>
<td>55</td>
</tr>
<tr>
<td>2.4 Past present and future satellites for the remote sensing of sea ice</td>
<td>56</td>
</tr>
<tr>
<td>2.5 Coverage, resolution and limitations of various sensors</td>
<td>57</td>
</tr>
<tr>
<td>2.6 Wavelength coverage of different AVHRR channels</td>
<td>59</td>
</tr>
<tr>
<td>2.7 Emissivity of different sea ice types at 19GHz</td>
<td>61</td>
</tr>
<tr>
<td>2.8 Capability of spaceborne sensors to provide sea ice parameters specified in table 2.3</td>
<td>73</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CHAPTER 3</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1 Possible gate triplets selected by the Seasat onboard processor</td>
<td>83</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CHAPTER 4</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.1 Fresnel reflection coefficients for first- and multi-year ice ice at various temperatures</td>
<td>95</td>
</tr>
<tr>
<td>4.2 Thickness and salinity of ice categories shown in figure 4.9.</td>
<td>112</td>
</tr>
<tr>
<td>4.3 Mean surface roughness parameters determined over new ice</td>
<td>117</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CHAPTER 7</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.1 AVHRR images for comparison with Geosat altimeter data</td>
<td>152</td>
</tr>
</tbody>
</table>
LIST OF TABLES (continued)

CHAPTER 9

9.1 Iceberg melt rates 207

APPENDIX A

A.1 Ice floe surface roughness measurements 236
A.2 Meltpond radius and distribution 238
1.0 Introduction

The aim of the work described in this thesis is to develop a new means of monitoring global sea ice cover using satellite radar altimetry. An additional aim is to extend the measurements of precise surface height using altimeters into sea ice covered areas. The potential applications of this work are widespread and can be broken down into the following areas:

(i) CLIMATOLOGY - the monitoring of changing sea ice cover to indicate long term changes in the global climate.

(ii) METEOROLOGY - contributing data on aspects of sea ice cover which affect regional and global meteorology.

(iii) SEA ICE MODELS - providing sea ice parameters needed for input as boundary conditions and for validation.

(iv) OPERATIONAL APPLICATIONS - sea ice data for support of shipping and other operational applications.

(v) POLAR OCEANOGRAPHY/GEOPHYSICS - extension of oceanographic and geophysical measurements made by satellite radar altimeters into sea ice covered areas.

In this chapter we discuss briefly each of these applications in turn.
1.1 Monitoring global climate change

1.1.1 Components of the climate system

The climate system is complex and consists of numerous components all of which interact through a series of complex physical processes. The main components of the climate system are shown in figure 1.1.

Figure 1.1 A schematic representation of the components of the atmosphere-ocean-ice-land climate system (adapted from Houghton and Morel [1984]).

The response times (t) of each component to climate change vary widely [Houghton and Morel, 1984]:

- Atmosphere - most variable component
  - lower atmosphere, t~weeks

- Oceans - large energy/chemical store
  - upper ocean, t~months-years
  - lower ocean, t~centuries

- Cryosphere - land ice, t~$10^2$-$10^4$ years
  - sea ice and snow, t~days-years

- Land surface - hydrological systems, t~days
  - biomass, t~weeks-years
Note that sea ice shows a response time much shorter than other components, such as the ocean.

1.1.2 Global climate models

The most powerful tools used to understand better the climate system are climate models. In addition to providing insights into the interactions of the various climate components, climate models may be used to understand changes which occur in the climate system and to simulate possible changes due to mankind's activities. The most complex type of models are 'Global Circulation Models' which use primitive equations governing the physical behavior of the atmosphere to compute the interaction between discrete elements over a series of time intervals. The elements form part of a grid which, for the most complex models, have horizontal sizes of a few hundred km and from two to more than ten vertical intervals. In general these models compute the behaviour of the atmosphere given a set of fixed boundary conditions such as radiation input, sea surface temperature, sea ice extent etc.. More recently, attempts have been made to produce coupled ocean/atmosphere models which allow both components to act as dynamic systems. This is important as the transport of heat, from equatorial to polar regions, by the ocean is at least as great, in mid-latitudes, as that transported by the atmosphere [Kellog, 1979]. In the future such models must also aim to include elements such as sea ice and snow cover as dynamic components.

1.1.3 Climate feedback mechanisms

The successful use of present day models depends on the observation of various climatic parameters both to allow boundary conditions to be specified and also to compare model predictions with real change. Future models will require a much better understanding of the physical processes governing interaction between the various components of the climate system. In addition important feedback mechanisms, which can act to amplify (positive feedback) or reduce (negative feedback) climatic change, must also be considered. Major feedback mechanisms related to temperature are listed below:

- Ice albedo (positive) - increasing $T >$ decreases ice cover $>$$\text{decreases albedo} >$ increases absorption $>$$\text{increases } T$
• Water vapour/radiation (positive) - increasing T > increases water vapour > increased radiation blanket > increases T

• Cloud/radiation (negative) - increasing T > increases cloud cover > decreases T

The effect of ice albedo feedback is thought to be especially important for climate monitoring as it results in any change in Global mean temperature being amplified in the polar regions.

1.1.4 The possibility of global climate change

There exists widespread evidence that the Earth's climate has undergone many substantial changes in the past. A hundred million years ago the dinosaurs inhabited a mainly tropical land with temperatures at high latitudes much greater than at present. In contrast, during the last glacial period, 20000 years ago, ice sheets covered most of Canada and Northern Europe. The larger fraction of water taken up in snow and ice meant that the sea level was some 80m lower than at present. More recently the most important climatic event in modern history is the so called 'little ice age' which occurred from the 14th to the 19th century. Since that time the indications are that a slight warming has occurred at least in the Northern hemisphere [Houghton and Morel, 1984].

In the past such changes are thought to have originated from external events, such as changes in the Earth's orbital parameters, and in the power output of the Sun. Sudden changes have been attributed to catastrophic events such as meteor impact and volcanic eruptions. At present, however, the greatest threat to the delicate balance of the Earth's climate comes from the activities of mankind.

Increases in the atmospheric levels of Carbon di-Oxide (CO₂) and other chemicals, such as nitrous oxide, methane and chloro-fluoro carbons, are the most significant in terms of mankind's influence on global climate Kellog [1979]. These gases allow incident solar energy, at visible wavelengths, to pass through the atmosphere but block the longer wavelength infra-red radiation re-emitted from the surface, causing a phenomenon commonly referred to as 'Greenhouse warming'.
Although it is difficult to predict future patterns in the production of CO$_2$ by man it is generally accepted that the atmospheric concentration of CO$_2$ will have double before the year 2100. Model estimates of the increase in mean global surface atmospheric temperature due to such a change vary, but the consensus is that a change of between 1 to 3 K can be expected. The ice/albedo feedback will serve to increase this change to between 5 to 10K in the polar regions. Figure 1.2 shows the best and worst case predictions for global and polar temperature rise due to greenhouse warming.

One possibility is that an increase in mean global temperature might lead to the complete disappearance of the present Arctic ice pack. Flohn [1982] presents paleo-climatic evidence that such a situation has existed in the past. Evidence from modelling work [Maykut and Understeiner, 1971] shows that a 10% change in the melting season could lead to the melting of the ice pack on a scale of 30-40 years. The increased absorption of radiation in the arctic ocean would amplify the temperature increase in northern latitudes. "The effects on global climate would likely be substantial, with the major climate zones shifting 2° or more northward", [Flohn, 1982].

The consequences of a significant increase in global temperature could be very serious. Results of a model simulation by Manabe and Wetherald at the University of Princeton show that the doubling of atmospheric CO$_2$ would turn the Great Plains in the United States, presently the source of a large fraction of the worlds grain, into a dry zone. The increase in temperature may also lead to significant melting, or even collapse, of the continental ice sheets. The disappearance of the Greenland or West Antarctic ice sheets would cause a 5-6m
rise in global sea levels, causing serious problems to many low lying regions, including capital cities and major industrial areas [Kellog, 1979].

It will be vital to monitor and understand better the Earths climate if human society wishes to prepare for such changes, and especially if the impact of mans' activities is to be controlled. The consequences of even minor changes in climate can be seen in the starvation caused by recent droughts in Africa and Asia and also in the effects that severe weather conditions can have, especially in more developed countries.

1.1.5 Monitoring Global climate change

The observations best suited to an early detection of global climate change have been listed as follows [Parker, 1985]:

(i) Atmosphere  - Surface, Tropospheric and Stratospheric temperature
    - Atmospheric Circulation
    - Precipitation
    - Radiation budget

(ii) Ocean  - Sea surface and deep ocean temperature
    - Ocean currents
    - Salinity
    - Global sea level

(iii) Cryosphere  - Snow cover
    - Sea ice extent

Parker notes that the greatest problems in detection of long term climate change using past observations arise from changes in instrumentation, observational procedures and locations. In principle satellite borne sensors can overcome most of these problems. They can also provide global sampling to take into account regional fluctuations which may mask real changes. In the polar regions, where the greatest increase in temperature is predicted, satellite data are particularly useful given the logistical problems in performing surface observations. Although the accuracies required for some of the parameters listed above are beyond the capability of present satellite sensors, the measurement of others is fairly well established. In particular the annual variation in sea ice cover, as monitored by satellites, has been studied by several researchers for detection of long term climate change.
Figure 1.3  Variations in interannual sea ice extent derived from satellite passive microwave observations [Gloerson and Campbell, 1988].

Figure 1.4  Predicted changes in mean Northern sea ice thickness for: (a) Present level of Carbon di-Oxide, (b) With a four fold increase in the present level of Carbon di-Oxide [Manabe and Stouffer, 1979].
1.1.6 Sea ice as an indicator of global climate change

The sensitivity of sea ice extent to climate change is not well known but empirical observations suggest that the change in sea ice extent could be as great as 2.5°lat/K in some regions of the Antarctic [Budd, 1975]. Although this is an upper estimate changes in the atmospheric temperature accompanying the predicted change in CO₂ concentration should cause a very large change in sea ice extent [Zwally et al., 1983b]. Observations of sea ice by Zwally et al. [1983b] and more recently by Gloerson and Campbell [1988], using satellite-borne passive microwave observations of sea ice show some significant downward trends in global sea ice cover. Figure 1.3 shows the total Arctic, Antarctic and Global sea ice extent derived using passive microwave data collected between 1978 and 1987 [Gloerson and Campbell, 1988].

Such a record is, however, considered too short to show a significant decrease in ice extent over the natural variability that occurs on time scales of 5-10 years. Model predictions also show that substantial changes in sea ice thickness could result from global warming induced by higher levels of CO₂ Manabe and Stouffer [1979]. Figure 1.4 shows their predictions of Arctic sea ice thickness for the present level of CO₂ and for a four fold increase in CO₂.

Apart from passive microwave instruments, the only other satellite-borne instrument presently capable of providing global synoptic monitoring, without the hindrance of darkness and cloud cover, of sea ice is the satellite radar altimeter. In this work we aim to explore the contribution that satellite radar altimetry might make towards the monitoring of global climate through observations of sea ice.

1.2 The role of sea ice in regional and global meteorology

Having discussed the passive role that monitoring sea ice might play in indicating long term climate change we now consider how sea ice affects regional and global meteorology.

1.2.1 Sea ice-Ocean-Atmosphere interaction

Sea ice plays an important role in controlling both regional and global variations in climate. The
important physical mechanisms can be summarised as follows:

(i) Insulation of the upper layer of the ocean blocking exchanges of heat, momentum and moisture between the ocean and the atmosphere.
(ii) Changes of upper ocean surface temperature through the absorption and dissipation of latent heat during melting and freezing
(iii) Absorption of solar energy leading to a reduction in surface albedo due to sea ice melting.
(iv) Changes of ocean salinity by freezing, brine drainage and melting of ice floes.

![Diagram of sea ice-ocean atmosphere interaction at the ice edge](adapted from Understeiner [1984])

The changes in local climate brought on by the presence of sea ice also affect the behaviour of sea ice itself. Important parameters and feedback mechanisms are shown in figure 1.5, although it should be stressed that the precise interaction between sea ice and climate anomalies is still not clear [Herman and Johnson, 1978].

1.2.2 Effects of sea ice on the atmosphere

The presence of sea ice reduces the heat flux from the ocean to the generally much colder atmosphere by as much as two orders of magnitude [Maykut and Understeiner, 1971; Allison, 1979; Gudmansen, 1983] thus reducing the turbulence of the lower atmosphere. This can act to change the track of polar depressions such as those which affect the weather in the United Kingdom. Where sea ice anomalies occur the resulting reduction in cyclone activity can lead to less heat advection from lower latitudes so further enlarging any
temperature anomaly.

Herman and Johnson [1978] used a global circulation model to predict variations in atmospheric pressure for observed minimum and maximum sea ice extents in the Arctic. Their results show significant differences in mean sea level pressure of 4 mb in the Davis straight and the Sea of Okhotsk and 8 mb in the North Atlantic between Iceland and the United Kingdom. Zonally averaged temperatures were observed to fall by 2°C below isobaric heights of 850 mb at latitudes from 50-70°N. A 13% increase in pole-ward energy transport between latitudes 40 and 53°N was also observed. Simmonds [1979] presents results from a Global circulation model of the Southern Hemisphere which show that differences in the March and September sea ice extent around Antarctica can cause changes in mean sea level pressure as far north as the equator. In addition changes in the sea ice cover may lead to changes in regional precipitation due to the blocking of ocean-atmosphere moisture exchanges.

1.2.3 Sea ice and global ocean circulation

The global transport of heat and salt by the world's oceans and the growth and decay of sea ice are intimately related.

In the Northern hemisphere interaction between sea ice cover and the global oceans is largely blocked by the presence of land masses. However, the role of ocean currents in controlling sea ice cover is clearly demonstrated in the European Arctic, with ice cover persisting far south on the Eastern coast of Greenland, due to the East Greenland current, whilst more northerly areas on the coast of Norway remain largely ice free due to the Gulf Stream (see section 2.2).

In the Southern Hemisphere sea ice does not come under the same constraints which occur in the Arctic. The Southern Ocean is therefore free to act as a giant heat sink for the world's oceans, with warm saline deep water from lower latitudes rising to release heat to the atmosphere.
Figure 1.6  Schematic diagram of deep water convection in the Southern ocean, adapted from Gordon and Comiso [1988].

Figure 1.6 shows a schematic diagram of the thermo-haline circulation in the Southern Ocean. At latitudes around 65°S warm saline water, from the equatorial regions, rises due to surface divergence of the ocean caused by prevailing wind conditions. Near the ice edge the rising warm water is cooled and its salinity is decreased, due to the greater precipitation, to form Antarctic surface and intermediate water. Further south, rising warm water can lead to polynyas (areas of open water within the ice pack), although low salinity and therefore low density meltwater can act to form a stable layer resulting in overturning of deep water below the surface. Near the coast, katabatic winds flowing off the Antarctic continent result in the formation of coastal leads where new ice is constantly formed. During the formation of new ice a large fraction of the salt in the sea water is expelled resulting in very saline cold water that falls to form Antarctic bottom water.

The mechanisms of deep ocean overturning and heat loss illustrate the important role that sea ice plays in controlling the transport of heat from the equator to the poles. In addition the mixing between deep and surface layers plays a major role in maintaining a balance between oceanic and atmospheric concentrations of gases such as CO₂ which influence global warming. The above discussion also shows that leads and areas of open water within the ice pack play a crucial role in controlling the transfer of heat from the ocean to the atmosphere at the poles. Studies of data from radar altimeters over sea ice carried out previously and herein show that they may provide a sensitive means of monitoring small areas of water which are
beyond the resolution of passive microwave instruments. The altimeter may provide valuable
data on lead fractions within sea ice on global scales.

1.2.3 Sea ice data important in determining influences on climate

In the previous sections we have briefly reviewed the processes governing ocean-atmosphere-sea ice interaction. Weeks [1981] summarises the sea ice parameters that are
important for different aspects of climate research:

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Influence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Albedo</td>
<td>Extent, type, snow cover</td>
</tr>
<tr>
<td>Insulation</td>
<td>Type, thickness, snow cover</td>
</tr>
<tr>
<td>Latent heat export</td>
<td>Thickness, drift velocity</td>
</tr>
<tr>
<td>Surface stress</td>
<td>Drift velocity, top and bottom ice roughness</td>
</tr>
<tr>
<td>Ocean mixed layer</td>
<td>Ice growth and ablation rates, drift velocity</td>
</tr>
</tbody>
</table>

1.3 Sea ice models

As with Global Climate models of the atmosphere, models of sea ice cover have been used to predict the future behaviour of sea ice cover under changing conditions. Most models developed so far concentrate on modelling sea ice alone, given boundary conditions for the atmosphere and the ocean. As stated previously, future aim is the production of coupled ocean-sea ice-atmosphere models.

![Processes controlling sea ice thickness.](image)

The evolution of sea ice after the initial growth stages depends on many factors. The main mechanisms which change the character of sea ice over a period of time are shown in figure
1.7. Some sea ice models consider only thermal processes whilst others also take into account dynamic effects. Hibler [1980] reviews three types of seasonal sea ice model:

(i) Global Thermodynamic models: Where sea ice growth and decay is allowed to vary over geographical grids driven by climatological data but ice motion is neglected.

(ii) Dynamic-thermodynamic models: Employing thermodynamic modelling similar to that of Washington et al. [1976] but also allowing ice motion.

(iii) Coupled ocean-atmosphere-sea ice models: Much simpler representation of sea ice but allowing for interaction between ocean and ice surfaces.

The low strength and high growth rate of thin ice mean that it dominates the thermal and dynamic properties of sea ice cover. Because of this many sea ice models (e.g. Parkinson and Washington [1979]) simply break down the ice thickness distribution into two components, thick ice and thin ice/open water. Maykut and Understeiner [1971] demonstrate that heat flux through thin (0-40cm) ice can approach that from the open ocean, i.e. one or two orders of magnitude greater than through thicker ice. Variations in sea ice thickness can significantly affect sea ice growth and decay. Thin ice and open water allow greater oceanic heating by radiation which can in turn increase bottom melting. For a stationary ice cover it has been demonstrated that, after the initial onset of decay and formation of meltponds, short wavelength radiation is primarily responsible for ice breakup and melting. Washington et al. [1976] show that changes of only a few percent in the lead fraction can have a significant effect on model predictions.

Hibler [1980] shows that by including dynamic effects, predictions of Arctic sea ice thickness distribution can be significantly improved. Empirical observations suggest that wind stress is the dominant component in the momentum equation with ice drift generally being 1-2% of the wind velocity. Current and tilt effects, although much smaller, may become important on longer timescales. Hibler identifies sea ice top and bottom roughness as particularly important in determining sea ice motion.

As part of the work carried out in the World Climate Research Program (WCRP), data requirements for inputs to and validation of sea ice models were identified. The basic requirements are given in table 1.1.
Table 1.1 Basic monitoring requirement for sea ice models [World Climate Research Program, 1983]

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Frequency</th>
<th>Resolution</th>
<th>Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ice concentration</td>
<td>~3d</td>
<td>30km</td>
<td>0-60 ± 5-10%</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>60-80 ±3-5%</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>&gt;80 ±2%</td>
</tr>
<tr>
<td>Ice thickness</td>
<td>-</td>
<td>-</td>
<td>0.5m</td>
</tr>
<tr>
<td>Ice velocity</td>
<td>~1d</td>
<td>500km</td>
<td>0.02ms⁻¹</td>
</tr>
</tbody>
</table>

To verify model results it is suggested that they be compared with the sea ice concentration, velocity and extent on a weekly basis. On annual and interannual timescales the model results should be compared with observed ice concentration and thickness budgets. It is also pointed out that although some successful comparisons have been made regarding ice in the Arctic basin far more data is needed to verify models covering the Antarctic.

Ice thickness is a key parameter which is needed for validation of many sea ice models. Thorndike [1980] attempted to validate model predictions of sea ice thickness but was hampered as lack of data prevented validation in particular of the distribution of thicker ice types. Zwally [1984] notes that that ice thickness is a primary variable in many sea ice models which are limited by lack of observational data.

1.4 Operational applications

1.4.1 Requirements for operational sea ice information

Sea ice presents a serious hazard both for shipping and oil exploration in the polar region. Kohler [1986] estimate that 231 ships sustained serious damage due to sea ice between 1970 and 1983 in an area from the Beaufort sea to the West coast of Greenland. Information regarding ice conditions is also needed for offshore drilling operations, where costs can run into many hundreds of thousands of pounds per day.
Information on sea ice conditions is generally required on three levels [Thompson, 1986]:

(i) Immediate - Information regarding sea ice conditions in the immediate vicinity of a vessel (0-20 km).

(ii) Tactical - Information for operations in the next few hours or days regarding ice conditions over an area of several thousand square km.

(iii) Strategic - Information as a general overview to be used in planning of routes or major operations required over areas of millions of square km.

Information regarding immediate ice conditions is normally obtained using ship borne radar or airborne support. Tactical information can be gained from high flying aircraft although such support is normally very expensive. Information from satellites can be of use in tactical support but minimising the time delay between observations and receipt of information at the point of use is critical. For strategic requirements satellite observations are the only practical means for satisfying operational needs outside areas where routine reconnaissance flights are made.

### 1.4.2 Use of satellite data for ice information and forecasting

Until the early 1970's the main source for data on ice conditions came from sparse shipborne observations and expensive aircraft reconnaissance [Lepparanta, 1986]. With the launch of the first NOAA satellites in 1970 it soon became clear the satellite remote sensing data could provide widespread coverage at relatively low cost. The operational use of satellite imagery involves data mainly from the Advanced Very High Resolution Radiometer (AVHRR) instrument which provides an average of twice daily coverage at medium (~1km) resolution imagery in visible and infra-red bands Mullane [1980]. High resolution visible imagery is sometimes used but coverage is much more restricted for Landsat being only 3 days out of every 17.

Early ice information systems used facsimile transmissions from satellite receiving stations but more recently direct digital links have considerably improved the quality of data available at ice centres [Ramsay et al., 1988]. Another recent development is the use of Automatic Picture Transmission (APT) receivers based on-board ships Brigham [1986].

35
1.4.3 Operational use of data from the Geosat altimeter

Visible and infra-red imagery is routinely used to provide ice maps for operational applications but their use is seriously affected in conditions of cloud cover. Since 1985 the Naval Operational Research and Development Activity (NORDA) have been generating information for use in operational forecasting using data from the Geosat satellite radar altimeter [Lybanon and Crout, 1987]. In later chapters we will explore how the current product, the 'Geosat Ice Index' (see section 2.4.5.9), can be improved.

1.4.4 Operational sea ice products from ERS-1

With the launch of the European ERS-1 satellite, in September 1990, it is planned to produce an operational sea ice product using data from the ERS-1 radar altimeter at the UK Earth Observation Data Centre (EODC). The design of algorithms to process the data has been already carried out but further research into the interpretation of the data will be needed to enable useful information for operational applications to be extracted [Laxon, 1989].

1.4.5 Sea ice parameters required for operational applications

Mulhane [1980] identifies the sea ice parameters important for operational applications as :

(i) Ice distribution - Ice concentration and ice boundaries
(ii) State of development - Discrimination between New, Young (10-30cm) and thicker first year ice.
(iii) Topography - metre scale ice roughness
(iv) Ice state - whether ice is forming or breaking up
(v) Ice movement - ice motion for input to forecast models

As mentioned previously a key requirement in providing sea ice monitoring for operational applications is that the time between observation and transmission to a user be minimised. Satellite radar altimeters provide global coverage at a relatively low data rate and normally require only simple ground processing. They are therefore well suited to some types of operational applications. In this work we will attempt to assess how radar altimeters can provide observations of the parameters listed above.
1.5 Polar oceanography and geophysics

1.5.1 Measurements provided by satellite radar altimeters

The use of satellite borne radar altimeters first began with the deployment of an instrument aboard Skylab in 1973. This was closely followed by instruments deployed on Geos-3 (1975) and on board the short lived Seasat mission, launched in 1978. More recently a dedicated mission was launched in 1985 aboard the US Navys' Geosat satellite which carried an instrument essentially identical to that flown on Seasat and which is still operational.

Satellite borne radar altimeters provide measurements of three geophysical quantities over the open ocean:

(i) Instantaneous sea surface height
(ii) Ocean waveheight
(iii) Windspeed

All instruments so far flown, and those planned for the near future, are designed specifically for operation over the open ocean. For technical reasons, which are discussed in chapter 3, the values for the above geophysical parameters are corrupted. In addition physical models which are assumed to govern the radar reflection observed by such an instrument are frequently invalid and a new approach to the interpretation of such data is required.

1.5.2 Altimeter measurements of sea surface height

The ocean gravity field varies in strength over the globe due to the presence of both topographic features on the ocean floor, and also due to the presence of more deep seated mass anomalies in the earths mantle. The marine geoid represents an equipotential surface in this field and deviates by as much as 100m from the so called 'reference ellipsoid'. If other effects such as currents and tides are ignored then the ocean surface can be said to represent such a surface. Hence measurements of sea surface topography directly reflect bathymetric features (sea mounts, ocean ridges and trenches) at short wavelengths and more deep seated anomalies (regions of upwelling and subduction) and longer wavelengths. Before the advent of satellite altimeters information on the global geoid was determined primarily from
satellite orbit data which could not resolve the shorter wavelength features associated with bottom topography.

The high accuracy and global coverage of satellite altimeter provides a unique opportunity to study the marine geoid. If data from a sufficiently long period is included in the analysis then transient variations due to tides and ocean currents can be greatly reduced, although this method will not eliminate permanent ocean currents. This data can be used to reveal many previously unknown bathymetric features, especially in the poorly surveyed regions of the southern ocean, e.g. [White et al., 1983; Sailor and Okal, 1983].

Where bathymetry data is available for comparison with that from a satellite altimeter the magnitude of the geoid response to bathymetric features can give a measure of the 'Flexural rigidity' or stiffness of the lithosphere, e.g. [Dixon et al., 1983]. The geoid response will tend to be less over younger seafloor where the effective elastic thickness is less and the lithosphere is weak and greater where the sea floor is older and thicker [Freedman and Parsons, 1986]. Such measurements will provide information on the age and thickness of the oceanic and continental lithosphere. The knowledge of these elements is the basis for understanding plate tectonics and earthquake mechanisms.

By comparing altimeter data with alternative geoid models and also by comparing near repeat tracks changes in sea surface elevation caused by ocean currents can be studied [Cheney and Marsh, 1981; Thompson et al., 1983]. If measurements of sea surface topography within sea ice areas could be made with sufficient precision this may allow mapping of surface currents in the important polar regions.

Figure 1.8 shows a mean sea surface map based on 18 days of data from the Seasat satellite by Marsh and Martin [1982]. Because of problems in operating over sea ice data below about 62°S and above 70°N have been excluded.

1.5.3 Altimeter measurements ocean waves

Estimates of metre scale surface roughness are made directly on board the satellite and are quantified by the Significant Wave Height (SWH), equal to four times the standard deviation of surface heights, assuming a Gaussian distribution [Mognard and Lago, 1979].
Figure 1.8: Global mean sea surface topography derived from Seasat altimeter data [Marsh and Martin, 1982].
Over the open ocean details of the sea floor bathymetry, such as fracture zones, seamounts and trenches are revealed. Data below about 60°S has been excluded due to contamination by sea ice.
Measurements of SWH by the Seasat altimeter have been shown to be in good agreement with data obtained from surface buoys by Fedor and Brown [1982]. Near the sea ice ocean boundary measurements of SWH can provide information on the penetration of long wavelength ocean waves (swell) into the ice pack, an important mechanism in breaking up ice floes [Rapley, 1984].

1.5.4 Extension of satellite radar altimeter measurements to the sea ice covered ocean

The measurement of mean sea surface elevation in areas of the ocean covered by sea ice has important applications, both in marine geoid mapping and also in sea surface dynamics. Data presented in chapter 9 will show that retrieval of significant geoid features can be achieved by correcting for instrumental problems in altimeter data over sea ice.

In the Northern hemisphere the coverage up to 82°N afforded by the ERS-1 satellite will allow the first opportunity to study the large part of the bathymetry of the Arctic basin [Anderson et al., 1988]. Although the sea floor bathymetry plays an important part in the Arctic basin ocean circulation it remains poorly surveyed due to the difficulty of ship operations in the high Arctic. From a geological point of view the Arctic basin is particularly interesting since it represents one of the youngest sea floors in the global oceans. Anderson has suggested that ERS-1 will provide coverage of geophysically important areas of the Arctic ocean. Results presented later will show, however, that he underestimates the problems that will occur in ERS-1 altimeter measurements of the marine geoid in the Arctic basin due to contamination by sea ice.

In the Southern hemisphere sea ice presents less of a problem for marine geoid mapping since most of the sea ice melts back during the Austral summer. Some regions, however, such as the Weddel and Bellinghausen seas, remain ice covered throughout the year. In retrieving accurate mean sea surface elevations under sea ice it may also be possible to extract information about ocean currents in important polar regions.
1.6 Summary

The motivation for pursuing the study of satellite radar altimetry over sea ice can be summarised as follows:

- The monitoring of climate change is vital if mankind is to prepare for future changes and assess the impact that man's activities might have. Changes will be most pronounced in the polar regions due to the ice-albedo feedback mechanism. Studies have shown that sea ice extent and thickness may be key indicators of climatic change.

- Sea ice plays an important role in influencing both the atmosphere and the oceans. Future coupled models will require information on sea ice cover, roughness and thickness which affect heat, momentum and moisture exchanges between the ocean and atmosphere.

- Sea ice models are key to understanding the behaviour of sea ice under changing climate conditions. Such models require estimates of surface roughness and lead fractions as input boundary conditions. Validation of sea ice models has been seriously hampered by the lack of data concerning sea ice thickness.

- Reliable, up to date, data on sea ice conditions is needed for operational applications such as ship routing and oil exploration. Key information is required on the location of the ice boundary, distributions of thick and thin ice, ice state (melting or freezing) and on ice motion.

- Satellite altimeters provide measurements of the marine geoid, ocean currents and ocean waveheight with numerous applications in the study of solid earth physics and ocean dynamics. The extension of these measurements into areas of sea ice covered ocean will provide additional measurements in potentially important areas.
The main aims of the work presented in this thesis can be summarised as follows:

(i) To gain a clear understanding of the operation of past and present space-borne altimeters over sea ice and the development of processing techniques to reduce the errors that occur.

(ii) To advance the understanding of the mechanisms governing the response of radar altimeters to sea ice surfaces.

(iii) To further the interpretation of satellite radar altimeter data over sea ice through comparisons of altimeter data with sea ice climatology and by comparison with other sensors.

(iv) To examine how data from satellite radar altimeters can complement observations from other sensors in contributing to applications outlined in this chapter.

(v) To explore the geophysical applications of satellite radar altimeter data which is currently available.
2.0 Introduction

In this chapter we discuss nature of the sea ice cover in the Arctic and Antarctic oceans and describe the physical characteristics of sea ice relevant to interpretation of microwave data. Previous work concerning the interpretation of data obtained from satellites over sea ice is then reviewed with attention focussed on previous analysis of radar altimeter data.

2.1 Physical characteristics of sea ice

Consideration of the physical characteristics of sea ice is necessary to determine electromagnetic properties important for remote sensing. Specific emissive and reflective properties of sea ice are considered more fully in section 2.4, concentrating on those for radar altimeters in chapter 4. Here we review how the evolution of sea ice will affect its overall physical characteristics.

The growth of sea ice starts with the formation of small (1-3mm) discoids of ice crystals. The subsequent evolution depends on the prevailing wind and wave conditions. In very calm conditions these discs would grow to form a very flat, thin layer of ice known as nilas, eventually developing into a consolidated sheet. However the turbulence in the upper layers of the water is normally such that the growth of the ice discs continues as small discs, known as frazil ice. In windy conditions this slurry will be herded downwind to form grease ice. In less windy conditions the ocean wave action will result in the formation of circular plates of ice with upturned edges, known as pancake ice (see Figure 2.1). Sea ice growth will continue in this manner with grease ice becoming trapped between the pancake ice, and eventually forming a consolidated cover [Squire, 1984].

The subsequent evolution of sea ice depends, to a large extent, on other factors, such as whether it lies in an area of compression or divergence, whether it undergoes melting and on the amount of precipitation. The overall pattern of ice evolution is shown in figure 2.2. New ice, with a thickness between 0-0.1m, will be the most saline, and will have a relatively smooth
Freezing Together and Rafting of Pancake Ice

Further Rafting and New Ice Formation

Figure 2.1 Formation of pancake and new ice [Lange, et al, 1989].

a) New ice

b) First-Year ice

c) Multiyear ice

d) Summer ice

Figure 2.2 Stages in the evolution of sea ice [Zwally, et al, 1983a].
surface. First year ice, with a thickness between 0.1-2m, may retain a fairly smooth surface, but with a small covering of snow on top. Multi-year ice (i.e. sea ice which has survived at least one melting season) will generally be more rough owing to greater weathering of the surface, and will have a lower salinity than first year ice due to drainage of brine (saline water which becomes trapped during ice formation). Meltponds may also appear on the surface of multi-year ice during periods of melt, sometimes re-freezing as the temperature drops back below zero.

![Figure 2.3 Ridging statistics for Arctic sea ice [Wadhams and Horne, 1980].](image)

The degree of ridging and thickness of multi-year ice will generally increase with age but will also depend on how much compression the ice cover has undergone. Few data have been collected concerning the large scale variations in sea ice thickness and ridging statistics. The best data so far collected comes from submarine sonar profiles. Some typical distributions for sea ice thickness are shown in figure 2.3 [Wadhams and Horne, 1980], they show a strong bi-modal distribution with a separation at an ice thickness of approximately 1.5m. The thinner class of ice consists of young first-year ice whereas the thicker ice is multi-year and ridged ice.
2.2 The nature of global sea ice cover

2.2.1 Differences in Arctic and Antarctic sea ice cover

Seasonal variation of sea ice extent differs considerably between the Arctic and Antarctic. In the Arctic the seasonal variation is estimated at 50% [Walsh and Johnson, 1979] with the minimum extent occurring in August and maximum occurring in February. In the Antarctic the seasonal variation is estimated at 75% with minimum extent occurring in February and maximum in August. Figures 2.4 and 2.5 shows the minimum and maximum extent of sea ice in the Arctic and Antarctic. The difference in seasonal variation of total sea ice extent arises largely from the landlocked nature of the Arctic basin which restricts sea ice growth during winter compared with the Antarctic where sea ice growth is unconstrained.

The Arctic ocean is almost completely enclosed by the continents of Asia and North America with exchange of polar and temperate water and sea ice masses occurring mainly through the gap between Greenland and Scandanavia. Figure 2.6a shows the prevailing oceanographic currents in the Northern hemisphere. The dominant oceanographic currents are the Beaufort gyre, which circulates around the North pole and the East Greenland and Norwegian currents.

Ice which forms in the high Arctic may remain in the Beaufort gyre for many years. Near Arctic coasts the circulation of the gyre results in shear zones creating large areas of deformed ridged ice. The main outlet for ice produced in the high Arctic is the southward flowing East Greenland current resulting in a very mixed ice regime on the east coast of that land mass. In contrast the Norwegian current carries warm Atlantic water northward into the region resulting in an ice free regime as far north as Spitsbergen for most of the year.

In the Southern hemisphere the growth and decay of sea ice is largely unconstrained by land. A much larger fraction of the sea ice formed during winter melts during the summer and multi-year ice is found in only a few regions. Sea ice in the Weddell sea tends to survive the summer due to the influence of the Weddell Gyre (see figure 2.6b).
Figure 2.4a Arctic minimum ice extent

Figure 2.4b Arctic maximum sea ice extent

Figure 2.4 Arctic minimum and maximum sea ice extent from passive microwave data [Zwally, et al, 1987a]. The ice boundary is defined to be 15% ice concentration.
Figure 2.5 Antarctic minimum and maximum sea ice extent from passive microwave data [Zwally, et al, 1983a]. The ice boundary at 15%, 50% and 85% is shown.
Figure 2.6a Arctic oceanographic currents [Zwally, et al, 1987a].

Figure 2.6b Antarctic oceanographic currents, adapted from Tchernia [1980].

Figure 2.6 Oceanographic currents in the polar regions.
(Location references in later chapters are also indicated.)
Differences in the structure and morphology of Arctic and Antarctic ice have also been observed. The greater precipitation in the Southern hemisphere causes some ice to sink under the weight of the snow cover resulting in a layer of frozen saturated snow (infiltration ice) on the surface. Little is known about the geometry of Antarctic ice although it has been established that undeformed first year sea ice is generally thicker (2.75-3.35m) compared with Arctic first year ice (2.0-2.2m). The percentage of deformed ice is also thought to be less in the Antarctic compared with the Arctic.

2.2.2 Zones Observed in Arctic seasonal sea ice cover

Surface observations that do exist show a highly dynamic and divergent ice cover in the seasonal ice zone. Wadhams [1980] describes three distinct zones in the seasonal sea ice cover:

2.2.2.1 Fast ice

Fast ice is generally the first to form in winter along the coast because of three factors: (i) The shallow water which reduces the depth of convection necessary to cool the water to freezing point, (ii) Lower water salinity due to river discharge, (iii) Generally calmer sea conditions. The ice forms as outlined previously and subsequent development depends on the shape of the shoreline, the bathymetry and the motion of the offshore pack. Landfast ice is stabilised either by pressure ridges which are grounded at various points, or by the restrictive geometry of constricted channels.

Pressure ridges arise due to shore-ward drives of the offshore pack which deform the fast ice cover. Grounding can also be caused when 'ice islands' become included in the fast ice cover. Eventually the fast ice cover will become thick enough to resist further deformation and will assume a permanent winter morphology. The distribution of pressure ridges is important due to the scouring of the sea bed which can threaten offshore pipelines and cables.

In constricted channels the date of break up of the fast ice cover in summer varies considerably from year to year, depending on prevailing climate and wind conditions. The monitoring of the
dates of break up of the fast ice has obvious applications to problems of navigation.

2.2.2.2 The Shear Zone

The shear zone is created when pack ice near the fast ice edge becomes heavily deformed by shear and convergence, generating a zone of heavier ridging than is observed further from the coast. Wadhams and Horne [1980] and Wadhams [1983a] present analysis of submarine profiles to obtain ridging statistics for sections of the Arctic ocean. The mean ice draft and percentage of thick ice are observed to decrease, approximately linearly, with increasing distance from the coast. The % undeformed ice also decreases steadily away from the coast. Although the mean draft of pressure ridges greater than 9m depth shows a negligible decrease moving away from the coast the frequency of such ridges decreases significantly. One interesting result from the observations is the lack of thin ice observed in the shear zone. This leaves unresolved the question of how the pressure ridges are formed. Wadhams suggests that, either thin ice exists for only a very short time with periods of compression in the pack of much larger duration than those of divergence or that ridging occurs throughout the shear zone or that pressure ridges are formed from normal thickness ice.

2.2.2.3 The Marginal Ice Zone

Several studies have been made of the character of sea ice in the Marginal Ice Zone (MIZ) of the Bering and Greenland seas. For example, the Bering sea MIZ shows three distinct zones from the ice edge inwards (see Figure 2.7 and Table 2.1).

The floe size distribution away from the ice edge depends mainly on the penetration of long wavelength ocean waves (swell) into the ice pack. Observations of swell penetration are discussed in more detail in section 4.7.1.1 in the context of their affects on radar altimeter observations.

The zonation occurring in the Greenland sea is not so well defined although a similar increase in floe size away from the ice edge is observed. Less rafting is seen in the edge zone and the small floes in this area are believed to have originated from the breakup of larger ice floes from within the pack [Squire, 1983].
Figure 2.7 Arctic sea ice zonation [Wadhams, 1980]. The size of ice floes generally increases away from the ice edge due to the attenuation of swell which causes large floes to break up near the ice edge.
Studies in both areas show that the zonation is strongly influenced by the local ocean wave climate in addition to other oceanographic and climatic factors.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Distance from ice edge</th>
<th>Floe size</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>(i) Edge zone</td>
<td>0 - 5 km</td>
<td>~10m</td>
<td>Waves causes buffeting of floes leading to ridged, uneven floes</td>
</tr>
<tr>
<td>(ii) Transition zone</td>
<td>5 - 30 km</td>
<td>50-100m</td>
<td>Wavefield attenuated resulting in smoother floes but floe size still limited by wave breakup</td>
</tr>
<tr>
<td>(iii) Interior zone</td>
<td>30 - km</td>
<td>≥100m</td>
<td>Wavefield mostly attenuated allowing much larger floes to develop</td>
</tr>
</tbody>
</table>

Table 2.1 Structure of the marginal ice zone of the Bering Sea [Wadhams, 1980].

2.2.3 Zones observed in Antarctic sea ice cover

Much less is known about sea ice conditions in the Antarctic, particularly during winter. Knowledge of winter ice conditions comes almost solely from coarse resolution passive microwave and infrared satellite data [Jacka et al., 1987]. Divergence is caused both by the prevailing northerly winds and the passage of atmospheric lows. The first 100 - 150 km of the ice pack shows a steady increase in floe size and thickness as the ice becomes more mature. However, during the late winter, throughout the MIZ newly formed thin ice represents a significant fraction of the total ice cover. Further into the ice pack the floe size is governed by the freezing together of smaller floes and brash ice rather than being limited by wave breakup.

Some observations do exist however and here we summarise zones observed from a cruise during June-September 1985 around longitudes 0° to 40° west [Wadhams, personal
Table 2.2 Zones within Antarctic sea ice cover

<table>
<thead>
<tr>
<th>Distance from ice edge</th>
<th>Sea ice character</th>
</tr>
</thead>
<tbody>
<tr>
<td>At ice edge</td>
<td>Fist sized pancakes with raised edges due to wave action</td>
</tr>
<tr>
<td>edge-100km</td>
<td>Independent pancakes increasing in size to 1-2m across with ~10cm surface roughness</td>
</tr>
<tr>
<td>100-400 km</td>
<td>Significant increase in floe size to 100's of metres eventually becoming consolidated with 5-10 km distances between areas of open water. Sea ice thicknesses between 50-80 cm with 20cm snow covering.</td>
</tr>
<tr>
<td>&gt;400km</td>
<td>Ridging of 1-3m observed with some multi-year floes observed at 67°S</td>
</tr>
</tbody>
</table>

Zonation of sea ice in the Antarctic is therefore similar to that observed in the Arctic, but on a larger scale.

2.3 Data requirements for sea ice monitoring

In chapter 1 we briefly reviewed the sea ice monitoring requirements for different applications in climate, glaciological and operational areas. Before examining the capabilities of various satellite sensors it is worthwhile reviewing the data requirements as specified by the ICEX (Ice and Climate EXperiment) working group.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Accuracy</th>
<th>Observational Requirement Space</th>
<th>Time</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boundary</td>
<td>5(20) km</td>
<td>5(20)km</td>
<td>1(3)d</td>
</tr>
<tr>
<td>Concentration</td>
<td>2(5)%</td>
<td>25km</td>
<td>1(3)d</td>
</tr>
<tr>
<td>Albedo</td>
<td>0.02(0.04)</td>
<td>25(100)km</td>
<td>1(3)d</td>
</tr>
<tr>
<td>Motion</td>
<td>0.1(1)km/d</td>
<td>5(100)km</td>
<td>1(7)d</td>
</tr>
<tr>
<td>Ridging:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Density</td>
<td>10(50)%</td>
<td>50(100)m</td>
<td>7(30)d</td>
</tr>
<tr>
<td>Orientation</td>
<td>10(30)°</td>
<td>n/a</td>
<td>7(30)d</td>
</tr>
<tr>
<td>Height</td>
<td>1(5)m</td>
<td>n/a</td>
<td>1(30)d</td>
</tr>
<tr>
<td>Ice type</td>
<td>5(10)%</td>
<td>1(25)km</td>
<td>7(30)d</td>
</tr>
<tr>
<td>Leads:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fractional area</td>
<td>10(50)%</td>
<td>50(100)m</td>
<td>1(3)d</td>
</tr>
<tr>
<td>Orientation</td>
<td>10(30)°</td>
<td>NA</td>
<td>1(3)d</td>
</tr>
<tr>
<td>Floe position</td>
<td>20(100)m</td>
<td>20(100)m</td>
<td>0.25(2)d</td>
</tr>
<tr>
<td>Surface melting</td>
<td>Wet/Dry</td>
<td>25km</td>
<td>1(3)d</td>
</tr>
<tr>
<td>Surface temp.</td>
<td>1(3)°K</td>
<td>25(100)km</td>
<td>1(3)d</td>
</tr>
<tr>
<td>Ice thickness</td>
<td>0.2(1)m</td>
<td>25(100)km</td>
<td>7(30)d</td>
</tr>
</tbody>
</table>

Table 2.3 Data requirements for sea ice monitoring as defined by the ICEX working group, adapted from NASA [1979]. Figures quoted are for the desired accuracy, spatial and temporal scales, with values in parentheses given as a minimum.

In the following sections we will examine how well the requirements listed in table 2.3 are satisfied by current sensors and techniques. The aim of this exercise is to determine whether satellite altimeter observations of sea ice can complement and supplement measurements of sea ice parameters supplied by other sensors.

### 2.4 Remote sensing of sea ice

The realm of polar ice is the least known and understood part of the Earths' surface [NASA, 1979]. Long term climate research requires large scale synoptic measurements of sea ice characteristics. In the inhospitable polar regions the difficulty of operations and consequent sparse nature of ground measurements leaves many questions unresolved. Satellite remote sensing provides the only practical means of monitoring sea ice on the spatial and temporal scales necessary to provide useful data for climate monitoring and model validation. The application of satellite remote sensing to the monitoring of sea ice has been reviewed by several authors [Zwally, 1984; Thomas, 1984; Gudmansen, 1983; Parkinson 1983; Robin et al., 1983; Weeks, 1981].
In chapter one we outlined the requirements for sea ice data for climate monitoring, sea ice models and operational applications. In this section we review the capabilities and limitations of satellite borne sensors used to monitor sea ice in some detail. This is necessary in order to identify both how radar altimeter observations of sea ice can complement observations by other sensors.

<table>
<thead>
<tr>
<th>Satellite</th>
<th>Operational</th>
<th>Instruments</th>
<th>Orbit inclination</th>
</tr>
</thead>
<tbody>
<tr>
<td>Landsat 1,2,3</td>
<td>1972-78</td>
<td>MSS</td>
<td>99°</td>
</tr>
<tr>
<td>Landsat 4,5</td>
<td>1978-present</td>
<td>MSS, TM</td>
<td>98°</td>
</tr>
<tr>
<td>Spot</td>
<td>1984-</td>
<td>Visible scanner</td>
<td>99°</td>
</tr>
<tr>
<td>Geos3</td>
<td>1973-8</td>
<td>ALT</td>
<td>108°</td>
</tr>
<tr>
<td>Seasat</td>
<td>Jun-Oct 1978</td>
<td>ALT, SAR, SCATT, SMMR</td>
<td>108°</td>
</tr>
<tr>
<td>NOAA series</td>
<td>1970-</td>
<td>AVHRR</td>
<td>98°</td>
</tr>
<tr>
<td>Nimbus 5</td>
<td>1973-6</td>
<td>ESMR</td>
<td>99°</td>
</tr>
<tr>
<td>Nimbus 7</td>
<td>1978-87</td>
<td>SMMR</td>
<td>99°</td>
</tr>
<tr>
<td>DSMP</td>
<td>1987-</td>
<td>SSM/I</td>
<td>99°</td>
</tr>
<tr>
<td>Geosat</td>
<td>1985-</td>
<td>ALT</td>
<td>108°</td>
</tr>
<tr>
<td>ERS-1</td>
<td>Sep 1990-</td>
<td>ALT, AMI, ATSR</td>
<td>98°</td>
</tr>
</tbody>
</table>

**Key to instrument definitions:**
- ALT: Radar Altimeter
- AMI: Active Microwave Instrument (Combined SAR and SCATT)
- ATSR: Along Track Scanning Radiometer
- AVHRR: Advanced Very High Resolution Radiometer
- ESMR: Electrically Scanning Microwave Radiometer
- HRIR: High Resolution Imaging Radiometer
- MSS: Multi Spectral Scanner
- SAR: Synthetic Aperture Radar
- SCATT: Microwave Scatterometer
- SMMR: Scanning Multi-channel Microwave Radiometer
- SSM/I: Special Sensor Microwave Imager
- TM: Thematic Mapper

**Table 2.4** Past, Present and Future Satellites for the remote sensing of Sea ice
Table 2.4 summarises the past, present and soon to be launched satellites which carry instruments suitable for providing information on sea ice.

Table 2.5 shows the typical coverage, resolution and limitations of the various instruments which are used for the remote sensing of sea ice. When it is considered that polar regions spend up to 6 months of the year in darkness and also that some areas, particularly near the ice/ocean boundary, suffer from a 80-90% cloud cover [Ackley, 1979], it can be seen that microwave instruments provide the only means for synoptic coverage.

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Swath width</th>
<th>Resolution</th>
<th>Limitations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Visible</td>
<td>200 km</td>
<td>10-30m</td>
<td>Cloud cover, Darkness</td>
</tr>
<tr>
<td>(Visible, MSS, TM)</td>
<td></td>
<td></td>
<td>Limited coverage</td>
</tr>
<tr>
<td>Infra-red</td>
<td>2500km</td>
<td>1-4 km</td>
<td>Cloud cover</td>
</tr>
<tr>
<td>(HRIR, VHRR, AVHRR, ATSR)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Passive microwave</td>
<td>1200km</td>
<td>30km</td>
<td>Low resolution</td>
</tr>
<tr>
<td>(ESMR, SMMR, SSM/I)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Active Imaging</td>
<td>100km</td>
<td>30m</td>
<td>Expensive processing</td>
</tr>
<tr>
<td>(SAR, AMI)</td>
<td></td>
<td></td>
<td>Limited coverage</td>
</tr>
<tr>
<td>Radar Altimeter</td>
<td>10km</td>
<td>700m (along track)</td>
<td>Poor spatial sampling</td>
</tr>
</tbody>
</table>

Table 2.5 Coverage, resolution and limitations of various satellite borne instruments

The swath width of an instrument dictates the revisit time/total coverage afforded by each sensor, depending on the orbit configuration. For example the passive microwave and infra-red instruments can provide complete global coverage several times a day. Visible sensors, such as Landsat, provide coverage of several days in succession, but followed by intervals of up to two weeks. Technical constraints limit the coverage afforded by SAR instruments such as the ERS-1 AMI. Power constraints limit operation of the AMI in imaging mode to 10% of the time and the high telemetry rate means that data can only be acquired when the satellite is in
Radar altimeters differ significantly from imaging sensors, providing only single point measurements along the sub-satellite ground track. Although the poor spatial sampling of satellite radar altimeters is frequently cited as a serious limitation in sea ice studies, the spatial sampling interval achieved over a 17 day period in the polar regions is similar to passive microwave instruments (see chapter 8).

2.4.1 Visible/Infra-red observations of sea ice

Visible imagery relies on the difference in albedo, the percentage of sunlight reflected from the surface, to distinguish sea ice from the surrounding ocean surface. The contrast between sea ice and ocean is greatest during the polar summer when the elevation of the sun is greatest.

Visible sensors onboard the Landsat and SPOT satellites can provide images of sufficiently high resolution (~10-80m) to allow delineation of individual floes and narrow leads. Visible data may also be used to determine floes size and shape and measure ice concentration. Ice type may also be determined through the changing albedo as thin ice grows to a thickness of one or two metres. Coverage is, however, very limited. For example, Landsat images cover an area only 185x185 km and the interval between revisit time can be as long as 17 days. Data is also not available when cloud cover is present and also during the long periods of darkness which occur during the polar winter.

Infra-red instruments provide measurements of emitted thermal radiation from the surface. The radiance or 'Brightness temperature', $T_b$, of the surface is the product of the physical temperature, $T$, and the emissivity, $\varepsilon$, of the surface thus:

$$T_b = \varepsilon.T$$  \hspace{1cm} 2.1

Both ocean and sea ice have similar emissivity values at infra-red wavelengths and the contrast between them therefore relies mainly on differences in surface temperature. The contrast between sea ice and ocean temperature is largest during winter when the lowest surface temperatures occur.
By using observations in both visible and infra-red wavelengths better discrimination of sea ice can be achieved. The Advanced Very High Resolution Radiometer (AVHRR) provides images in five visible and infra-red channels at a resolution of 1km.

<table>
<thead>
<tr>
<th>AVHRR channel</th>
<th>Band</th>
<th>Wavelength range</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Visible</td>
<td>0.58-0.68µm</td>
</tr>
<tr>
<td>2</td>
<td>Visible/Near Infra-red</td>
<td>0.73-1.10µm</td>
</tr>
<tr>
<td>3</td>
<td>Reflected Infra-red</td>
<td>3.55-3.93µm</td>
</tr>
<tr>
<td>4</td>
<td>Thermal Infra-red</td>
<td>10.5-11.5µm</td>
</tr>
<tr>
<td>5</td>
<td>Thermal Infra-red</td>
<td>11.5-12.5µm</td>
</tr>
</tbody>
</table>

Table 2.6  Wavelength coverage of different AVHRR channels

Condal et al. [1985] compared AVHRR and ice chart data to show that different thin ice classes could be distinguished using a linear combination of Channels 1 and 2. This analysis relied on results from a model developed by Grenfell [1983], showing that the Albedo of sea ice increases gradually up to a thickness of about 1m, beyond which changes are minimal. Such analysis can, however, be thwarted by snow cover, which reduces the contrast between the different ice types and also in areas of inhomogeneous ice cover, where both ice and water appear in a single image pixel.

LeSchack [1974] investigated the potential of infra-red observations at longer wavelengths for ice thickness mapping, using data from the earlier NOAA-1 satellite with similar spatial operating at a wavelength (10.5-12.5µm). As the thickness of sea ice increases the surface observed by the satellite becomes insulated from the warmer water beneath. At some point the ice will become sufficiently thick that its surface temperature will be similar to the air temperature which, during winter, will typically be several degrees below the ambient water temperature (-2°C). Differences in surface temperature allowed discrimination of first year ice (thickness ~ 1m) and multi-year ice (thickness ~2-3m). Comparison of the concentration of the different ice types with data collected in the field showed good agreement. The technique is, however, limited since absorption by atmospheric water vapour can attenuate the radiation emitted by the sea ice. In addition snow cover can also affect surface temperature due to its low thermal conductivity.
The provision of two infra-red channels, with the AVHRR instrument, allows some correction for atmospheric absorption but cloud cover still remains a serious problem [Wannamaker, 1984]. Data from channel 3 of the AVHRR is best suited to cloud/ice discrimination but unfortunately suffers from instrumental noise which has hampered its use.

Data from the AVHRR instrument is widely used in operational applications [Condal and Le, 1984; Condal et al., 1985; Ramsay and Zieger, 1986] but poor knowledge of the physical relationship between observed brightness temperature and variable ice properties inhibits quantitative measurements. Cloud cover remains a serious problem and quoted accuracies for ice concentration estimates from AVHRR are as poor as ±25% [Cavalleri et al. 1983]. Analysis therefore relies mainly on image processing techniques (contrast stretching, textural analysis and principal components analysis) and human interpretation.

2.4.2 Passive microwave observations of sea ice

The use of passive microwave data from both airborne and satellite platforms in monitoring sea ice is now well established. Passive microwave data has been used to monitor the seasonal and inter-annual variation in sea ice cover that may reveal changes in global surface temperatures (see section 1.2.6) [Zwally et al., 1983b; Gloerson and Campbell, 1988]. The most comprehensive presentation of passive microwave data over sea ice is given in two atlases by Zwally et al. [1983a;1987a] using data from the Electrically Scanning Microwave Radiometer (ESMR). The ESRM was flown on the Nimbus 5 satellite between 1973 and 1976 and provided observations at 19GHz with a spatial resolution between 25-35km.

Passive microwave instruments measure the 'Brightness temperature' at microwave frequencies but instead of discriminating ocean from sea ice due to temperature contrasts, rely on the difference in emissivity. The emissivity of different sea ice types, at a frequency of 19 GHz, are shown in table 2.7. Estimates of ice concentration are made by first determining brightness temperatures of ice free and 100% ice covered regions. Linear interpolation can then be used to derive ice concentration to an accuracy approaching ±5% [Zwally,et al. 1983a]. However variation in the water emissivity, caused by changing sea state, means that the ice boundary cannot be positively identified below 15% ice concentration.
Table 2.7 Emissivity of different sea ice types at 19GHz [Zwally et al., 1983a].

<table>
<thead>
<tr>
<th>Ice type</th>
<th>Emissivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>0.5</td>
</tr>
<tr>
<td>New ice (&lt;10cm)</td>
<td>0.45 - 0.92</td>
</tr>
<tr>
<td>First year ice</td>
<td>0.92</td>
</tr>
<tr>
<td>Multi-year ice</td>
<td>0.84</td>
</tr>
<tr>
<td>Summer ice</td>
<td>0.45-0.95</td>
</tr>
</tbody>
</table>

The difference in emissivity between first-year and multi-year ice requires regions of multi-year ice to be identified so that the correct emissivity can be assumed to compute ice concentration. Zwally et al. [1983a] used observations during the minimum extent to identify areas where sea ice survived during the summer season. Variations in the emissivity of multi-year ice can, however, cause problems. Carsey [1982] examined data from the ESMR near the onset of fall, when the percentage of first year ice is essentially zero, to look at possible temporal and spatial variations in the emissivity characteristics of multi-year ice, possibly caused by scattering from ice ridges and meltponds. He identified significant variations, of 3-4%, in the multi-year ice emissivity which, if employed to map the lead fraction within the Arctic pack ice would lead to a possible 100% error in the estimates of total heat flux.

The launch of the Scanning Multi-channel Microwave Radiometer (SMMR) on board Nimbus-7 allowed measurements at five different frequencies (6.6, 10.7, 18, 21, and 37GHz) in both vertical and horizontal polarisations. The instrument resolution is frequency dependent ranging from 30km at 37GHz to 150km at 6GHz. Since the emissivity of different sea ice types varies with frequency data from the SMMR may be used to discriminate between different ice types by combining different microwave channels.

Svendsen et al. [1983] used a combination of the 10 and 37GHz vertical polarisation channels, giving a resolution of 90 km. By comparing with airborne observations they estimate the accuracy of total ice concentration measurements to be ±3% and multi-year ice concentration to be ±10%. A combination of the 18 and 37GHz vertical polarisation channels gave an improved resolution of 60km, but with slightly less accurate concentration measurements. An accuracy for ice edge location of ±10km for the SMMR was quoted, although the Norwegian Meteorological institute observed a 40km standard deviation in
SMMR ice edge locations compared to those determined using AVHRR [M., Kristensen, personal communication].

Passive microwave instruments have provided global synoptic data on sea ice extent and areal coverage which are needed both for climate monitoring and for global climate and sea ice models. The most serious limitations to the use of passive microwave observations are the low resolution, and the uncertainty about the microwave emissivities of the water and sea ice surfaces. This impacts on mapping of the sea ice boundary, particularly during the melt season when the contrast between ice floes and the ocean is reduced. This is clearly an area where data from satellite altimeters may be able to contribute as sensitive detectors of the sea ice boundary at a resolution which is potentially much better than passive microwave instruments. In a later chapter a comparison of the inter-annual variation in sea ice extent mapped by radar altimetry and by passive microwave instruments is presented.

2.3.4 Synthetic Aperture Radar observations of sea ice

Synthetic Aperture Radar (SAR) provides high (~30m from satellites) resolution imagery that is unaffected by darkness or cloud cover. Since the radiation source is the instrument itself, phase information in the return signal allows the reconstruction of the surface at much higher resolution than would otherwise be possible. Because of this potential SAR is widely regarded as the prime instrument for sea ice monitoring and much effort has been expended in its interpretation with a view to analysing data from the ERS-1 Active Microwave Instrument (AMI). In particular techniques for the automatic extraction of sea ice parameters from SAR data have been developed to cope with the large data volumes expected from future instruments.

Discrimination between different ice types may be achieved through changes in the observed radar backscatter, caused by variations in surface roughness and di-electric properties. Generally the off-nadir backscatter will increase with increasing surface roughness and salinity. Thus rougher multi-year ice will normally have a higher backscatter than first-year ice, whilst both will have a larger backscatter than smooth water lying between floes.

One important characteristic of data from Synthetic Aperture Radars arising from the coherent nature of the radiation observed is so called 'fading noise' which arises from the changing interference between reflecting facets on the surface as the instrument moves. Fading noise
is usually reduced by averaging data from several individual 'look's at the surface, but is still large enough to prohibit the use 'pixel intensity' type classification schemes employed on other types of imaging data. Discrimination of different sea ice types using SAR data has been carried out using statistical measures such as the mean, standard deviation, skewness and kurtosis of pixels within a given sample area [Lyden et al.,1984]. More complex measures such as entropy and inertia have also been employed to distinguish different ice types [Holmes et al.,1984]. The high resolution of SAR also allows measurements of floe size distribution and ice motion [Burns et al., 1985]. Ice concentration can also be determined by identifying areas of 100% water and ice cover and employing linear interpolation techniques.

Most analysis carried out so far has used using airborne observations. Spaceborne SAR observations have only been made by the short lived Seasat satellite and also using the Shuttle Imaging Radar. The use of Seasat SAR data over sea ice has primarily been in studies of ice motion since the radar parameters (1.9GHz, 23° incidence angle) were not ideal for ice type discrimination [Onstott et al., 1982].

Although much work has been undertaken to measure radar backscatter from sea ice, at different frequencies, incidence angles and polarisations, variation in radar backscatter from sea ice is still not fully understood. Small changes in surface conditions can result in significant changes in the backscatter from a particular ice type. Wind can change the backscatter from open water areas within a short time and surface wetness complicates interpretation. It should be said that this problem affects all observations of sea ice using radar, including radar altimetry. In chapter 4 we will consider the problem of radar backscatter from sea ice in more detail. Although studies over small areas have shown some potential for automatic classification of sea ice its seems unlikely that these will be universally applicable.

Much effort is being expended in preparing to analyse data from the ERS-1 AMI instrument. Campaigns such as BEPERS (Bothnian Experiment in Preparation for ERS-1) have endeavoured to improve interpretation of SAR imagery through airborne and field measurements [Thompson and Lepperanta, 1987]. Although the ERS-1 AMI operates only 10% of the time, due to power constraints, it will generate ~10 GBytes of data per orbit. The analysis of such large volumes of data from such instruments presents serious practical problems. It seems unlikely that SAR will be able to provide observations of sea ice on a global
scale in the near future.

2.4.4 Radar altimetry over sea ice

The potential use of radar altimeter data in sea ice monitoring has been mentioned in several references [Thomas, 1984; Gudmansen, 1983; Parkinson 1983; Robin et al., 1983; Robin, 1984; Weeks, 1981; Squire et al., 1984].

2.4.4.1 Analysis of GEOS-3 altimeter data

The first real attempt to quantify the response of satellite radar altimeters over sea ice were carried out by Dwyer and Godin [1980] using analysis of data from the GEOS-3 altimeter. They noted both an increase in return power, and a change in return waveform shape as the altimeter passed from open ocean to ice covered surfaces. They determined that waveform shapes over sea ice were related to specular surface reflections, with a sharp fall off in return power as the altimeter sampled areas further from normal incidence, in contrast to returns over the ocean which are essentially isotropic in the narrow range of incidence angle sampled by the altimeter. In an attempt to quantify the altimeter return signal they developed an 'Ice Index' which is calculated using:

\[
\text{Ice index}_{\text{GEOS-3}} = \left[ \frac{100+\text{AGC}}{100^\text{ASG}} \right] - 10
\]

2.3

The AGC (Automatic Gain Control) value represents an measure of the strength of the return echo and the ASG (Attitude Specular Gate) is a measure of the degree to which the return power falls off with increasing incidence angle. Thus over sea ice the value of AGC will tend to increase whilst the value of ASG will decrease. (Details of altimeter operation are more fully covered in chapter 3).

The primary aim of the 'Ice Index' was the detection of transitions from open water to sea ice using a threshold on the value of the ice index. By comparison with ice boundaries, determined from published ice charts, the Ice Index was formulated to be negative over the open ocean and positive over sea ice, the zero crossing point indicating an ocean/sea ice boundary. It was also suggested that variability observed in the 'Ice Index' value might be related to changes in ice concentration and ice type. By comparing the 'Ice Index' with ice types identified in ice charts they concluded that lower positive values of Ice Index
corresponded to low ice concentration or a high concentration of rough ice. Higher values corresponded to smoother younger ice.

2.4.4.2 Ice freeboard measurements using satellite radar altimetry

The potential of space borne altimeters to provide direct measurements of ice freeboard was investigated by Stanley et al. [1980]. They analysed altimeter derived elevation measurements, also using GEOS-3 data, for collinear tracks corresponding to different seasonal sea ice extent. The data analysed was collected in March 1976 and May 1978 in the region of the Bering sea.

![Figure 2.8](image_url)

**Figure 2.8** Elevation profiles over sea ice with sea ice present (Mar) and absent (May) [Stanley et al., 1980]. Although a significant difference in elevation is observed this can be attributed mainly to instrumental errors which occur when the altimeter operates over sea ice.

By comparing elevation data in areas where sea ice was present in one data set and absent in the other, they hoped to discriminate between signals caused by geoid undulations and those caused by the satellite measuring to the top of ice floes rather than to the ocean surface. Their results, reproduced in figure 2.6, show a significant, one to two metre, increase in elevation with sea ice present. Although the results seem encouraging at first it was stated that instrumental problems which occur when operating over sea ice could be responsible for an error of up to 1.8m. Another reason to question their results is that the ice freeboard
necessary to cause the differences they observe implies an ice thickness of around 10-20m whilst examination of ice maps provided by Zwally et al. [1987] show ice in this area to be seasonal and is therefore unlikely to have a thickness greater than a few metres.

2.4.4.3 Satellite radar altimeter observations of swell penetration

Profiles of significant wave height (SWH) measured by the Seasat altimeter we analysed by Rapley [1984] near the ice boundary to determine whether swell penetration (see 2.2.2) could be measured using radar altimetry. The distance between the ice boundary, determined using the AGC value and the point at which the satellite measurement of SWH fell near zero was taken as a measure of swell penetration. Analysis for the whole southern ocean showed correlations between swell storms in the open ocean values for SWH and significant swell penetration into the pack.

2.4.4.4 Physical interpretation of radar altimeter return waveform shapes over sea ice

Robin et al. [1983] consider physical models which may be used to explain radar altimeter return waveform profiles. These are:

(i) Plane polished reflector
(ii) Perfect diffuse reflector
(iii) Extended rough surfaces

Physical models for these three types of surface are described in more detail in chapter 4. They show that for some extended rough surfaces (with maximum slopes of $\sim 3 \times 10^{-3}$), the return power from a small fraction ($0.01\%$) of the surface may dominate the return waveform. In addition it is shown that such surfaces may result in waveforms which closely resemble that which would results from a plane polished reflector.

In interpreting waveform data from the Seasat altimeter obtained over sea ice Robin et al. [1983] suggest that damping of ocean waves by sea ice is an important mechanism in reducing the overall slope of the scattering surface although the lack of surface slope observations in sea ice areas is acknowledged to be a problem.

In a later paper Robin [1984] suggest that interpretation of radar altimeter data over sea ice
would benefit from comparison of line profile information from radar altimeters with imaging instruments. This type of analysis has since been carried out using airborne altimeter and photographic data (Sections 2.4.4.5, 2.4.4.7) and near co-incident satellite SAR and altimeter data (Section 2.4.4.6). In later chapters of this thesis we present comparisons of radar altimeter observations made by the Geosat satellite with visible and infra red imagery from NOAA satellites.

2.4.4.5 Radar altimeter observations during MIZEX '84

During the Marginal Ice Zone EXperiment [McIntyre et al., 1987] an airborne altimeter was also deployed over sea ice providing an opportunity for direct comparison with high resolution visible imagery and surface observations.

Drinkwater [1987] employs four parameters to describe the waveform profile observed by the RAL (Rutherford Appleton Laboratory) airborne altimeter. Two measures of pulse shape are used. The RMS roughness represents the standard deviation of the return of individual pulses during construction of the mean waveform profile. The leading edge slope is computed by fitting a straight line to the five data points centred on the half power point of the leading edge. Two backscatter coefficients are calculated, the peak waveform backscatter appearing in the waveform and the mean backscatter value throughout the waveform.

Observations presented show high backscatter returns over both open water and areas of small floes with diffuse, 'ocean like', returns occurring over floes significantly larger than the altimeter footprint.

Drinkwater used an empirical model for sea ice backscatter suggested by Onstott [1980] and Kim [1984] which gives the backscatter \( \sigma^0(\theta) \) as :

\[
\sigma^0(\theta) = \sigma^0(0^\circ) - \left[ \frac{4.34 \theta}{\theta_0} \right] \text{dB}
\]

where \( \sigma^0(0^\circ) \) is the backscatter at normal incidence and \( \theta_0 \) is some constant representing the rate of fall off with incidence angle. Drinkwater fits this model to return waveforms obtained
over 20 large floes and observes four distinct classes with values for $\sigma^0(0^\circ)$ ranging from $-9$ to $-13$ dB and values for $\theta_0$ between 2.7 and 11.7°. Higher values of $\sigma^0(0^\circ)$ and $\theta_0$ corresponded to observations of floes with a lower rms roughness or less snow cover. Such results demonstrate a potential for discrimination of ice type for floes larger than the altimeter footprint for airborne systems. The sampling angles of space-borne systems is however much narrower (0.3° for Seasat/Geosat; 0.3-1.2° for ERS-1) than for the RAL altimeter (~5°) and the footprint, which must be filled by the ice floe to eliminate contamination by other types of surface, much larger (5 km for Seasat/Geosat; 5-20km for ERS-1). In chapter 7 we will show that diffuse low power returns are indeed observed in satellite altimeter data over vast ice floes.

Over open water areas Drinkwater employs the specular point theory which has been previously used by other researchers to derive a relationship between mean square surface slope and altimeter backscatter over the ocean. His fit of observations to such a model show reasonable agreement between 2-7° incidence angles although it is poor at angles less than 2°. Further discussion and a more detailed description of the specular point theory is left until chapter 4.

In the last section of his analysis of airborne altimeter data over sea ice, Drinkwater suggests an inverse relationship between ice concentration and the integral of waveform power which observations seem to support. Conservation of energy supports this argument if returns from calm water between ice floes dominates the altimeter return and if most of the energy returned from the surface appears in the range window. Under such circumstances the total integrated return power in the waveform will depend only on the Fresnel reflection co-efficient of the water, which is essentially constant, and the fractional area coverage of water. The assumption that most of the return power is returned to the altimeter may break down in two circumstances. Firstly small scale roughness where radar energy is scattered through Bragg scattering which is undetectable by the altimeter will act to simply decrease the overall power in the the profile whilst not altering it's shape (section 4.5.4). Another serious problem with spaceborne observations is the limited angular sampling which means that a large fraction of the return power in the waveform may appear only for the most peaked waveforms. Where suitable models can be applied to the data to essentially extrapolate the power outside the range window some progress may be made. This approach will be explored later using data from the Seasat altimeter (section 6.2).
2.4.4.6 Comparison of Seasat radar altimeter and Synthetic Aperture Radar observations of sea ice

Ulander [1987a; 1987b] compared data from the Seasat altimeter over sea ice with that obtained by the Seasat SAR. The Seasat SAR images at some 23° off nadir so that observations exactly coincident with the Seasat altimeter are not possible. Near the latitudinal limit however overlap does occur between consecutive orbits permitting comparison within a few hours. Ulander used this fact to compare SAR data obtained during one orbit with altimeter data from the two subsequent orbits, during October 1978 in the area of the Beaufort sea.

![Graph showing zones observed in Seasat radar altimetry compared with ice type identified in Seasat SAR imagery [Ulander, 1987b]. Letter codes show the different types of ice identified in the SAR imagery (OW = Open Water, NI = New Ice, MY = Multi-Year ice). The number code is used merely to identify the different zones along the transect. Parameters shown are $\sigma^0_1$ (solid line) and $\sigma^0_2$ (dashed line).]

Ulander employed three parameters to characterise the return waveform power and shape. Two measures of the radar backscatter were defined; $\sigma^0_1$ - a measure of the peak backscatter and $\sigma^0_2$ a measure of the backscatter in the trailing edge of the waveform. Figure 2.9 shows the parameter values for a transect over sea ice with different ice types identified in the SAR imagery. Both parameters show clear delineation of open water, new ice and more mature floes. The distinction between multi- and first-year ice was less clear although Ulander claims that backscatter from the first year ice is slightly higher. Comparisons of waveform profiles were also made against various backscatter models which are normally applied over the ocean or have been proposed for sea ice. These models are described in chapter 4 and a discussion of Ulanders model fitting appears in the discussion section of that chapter.
More recently Ulander [1988] suggests that different ice types can be distinguished using the peak backscatter value. A comparison between Geosat altimeter data and ice charts shows separation between open water ($\sigma^0 < 20\text{dB}$), new thin ice ($\sigma^0 \sim 32.5\text{dB}$) and fast ice ($\sigma^0 \sim 37.8\text{dB}$).

2.4.4.7 Comparison of Seasat altimeter data with ice charts

An investigation of the response of the Seasat altimeter to ice covered ocean was carried out by Fedor and Walsh [1988]. They compared altimeter data collected in the Beaufort sea during October 1978 with ice charts from from the Canadian Atmospheric Environment Service. Three dimensional plots of waveform profiles along with time line plots of the peak backscatter observed were compared with ice types identified in the charts. They noted a high level of peak backscatter, $\sigma_{\text{max}}^0 \sim 40\text{dB}$, over new ice reducing to 25-30 dB over multi-year ice. Other features were identified as open water, not identified in the ice chart and also an ice island which the altimeter traversed over a distance of around 5km.

They also formulated an empirical ice index, $I$, to detect the presence of sea ice and to possibly identify different ice types:

$$I = 100 \frac{t_p}{t_s} \frac{S_p}{P} \sigma_{\text{max}}^0$$

where $P$ is the mean power in the first five waveform samples after the leading edge and $S_p$ is their standard deviation, $t_p$ is the time constant of power fall off due to the antenna pattern, $t_s$ is the time constant of power fall off due to the specular nature of the surface and $\sigma_{\text{max}}^0$ is the peak backscatter. Details were not provided on the method for estimating the last two parameters in this list. Comparison of the ice index with the ice chart showed similar behaviour to the peak backscatter although with a more 'noisy' appearance.

2.4.4.8 Comparison of airborne altimeter data with visible imagery

Prior to the Seasat mission, in March 1978, an instrument similar to the Seasat altimeter was flown over the Beaufort sea. Fedor et al. [1988] compared results from the altimeter with co-incident high resolution aerial photography. Again the altimeter showed very high returns
over new ice and nilas with lower intensity returns over first year ice further decreasing over areas of ridging.

2.4.4.9 Operational use of radar altimeter data over sea ice

With the launch of the US Navy's Geosat satellite in 1985 it was decided that an operational sea ice product should be produced to complement visible imagery and other data used by the Joint Ice Centre in Suitland [Lybanon and Crout, 1987]. The Geosat ice index is a modified form of that developed by Dwyer and Godin (section 2.4.4.1) and is computed using:

\[
\text{Ice Index}_{\text{GEOSAT}} = \frac{100 + \text{AGC}}{100 \times \text{VATT}}
\]

Where VATT is computed using the normalised power in the last 8 gates of the return waveform thus:

\[
\text{VATT} = \frac{\frac{1}{8} \sum_{i=0}^{8} P_i - \sum_{i=1}^{8} P_i}{\frac{1}{48} \sum_{i=0}^{48} P_i - \sum_{i=1}^{8} P_i}
\]

The mean power in the first 8 gates of the return is subtracted to account for instrument noise.

Comparisons of the Geosat ice index with co-incident data from the AVHRR showed a high sensitivity to the location of the sea ice boundary. An error in the processing algorithm for the Geosat ice index, however, resulted in a noisy appearance and in numerous data gaps. An investigation into the cause of this problem (carried out by the author) revealed that the value of VATT recorded in the Sensor Data Record was not the same as that computed from the waveform profile.

During processing the telemetred value of VATT is converted into the value written to the SDR (VATT_{SDR}) using:
\[ VATT_{SDR} = a_n + b_n VATT_{TELEMETRED} \]

where \( a_n \) and \( b_n \) are constants which depend on the gate triplet selected by the onboard processor for waveform tracking (section 3.3.3). This problem has since been corrected and ice index values now show a less noisy appearance with fewer data gaps although sensitivity is apparently reduced [Hawkins and Lybanon, 1989].

2.4.4.10 Previous work on radar altimetry over sea ice - discussion

Previous work analysing altimeter data over sea ice has demonstrated the potential of such an instrument in monitoring sea ice. However all such studies have been essentially qualitative reflecting the lack of understanding concerning physical causes of variations in the altimeter signal over sea ice. Most previous studies also suffer from lack of ground truth and some have relied on ice charts of questionable quality. Of the work reviewed here only Ulander, Fedor et al. and Drinkwater use co-incident observations from other sensors. The studies by Fedor et al. and Drinkwater were carried out using airborne altimeter data which may differ considerably from spaceborne altimeters. In particular airborne systems generally sample over a broader beam width to compensate for poorer instrument pointing. In addition airborne systems may suffer from the near field effect (see section 4.4). In other studies instrumental errors have seriously compromised data interpretation (for example Stanley et al. [1980]).

In this thesis we will pursue the investigation of the altimeter response to sea ice surfaces through consideration of backscattering properties, examination of global patterns in the altimeter signal and by comparison with satellite imagery. The retrieval of elevation measurements over sea ice for geoid retrieval and to allow further examination of the possibility of freeboard measurement will also be pursued.
2.5 Satellite instrument capability for supplying sea ice parameters

The capability of satellite borne instruments to satisfy the requirements laid out in Table 2.3 is summarised in Table 2.8.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Visible/Infra red</th>
<th>Passive microwave</th>
<th>SAR</th>
<th>ALT</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boundary</td>
<td>Yes(±1km)</td>
<td>Yes(±30km)</td>
<td>Yes(±1km)</td>
<td>Yes(±1km)</td>
</tr>
<tr>
<td>Concentration</td>
<td>Yes (±25%)</td>
<td>Yes (±15%)</td>
<td>Potential</td>
<td>Potential</td>
</tr>
<tr>
<td>Albedo</td>
<td>Yes</td>
<td>No</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>Motion</td>
<td>Yes</td>
<td>Large scale only</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>Ridging:</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-Density</td>
<td>Possible</td>
<td>No</td>
<td>Possible</td>
<td>Potential</td>
</tr>
<tr>
<td>-Orientation</td>
<td>Possible</td>
<td>No</td>
<td>Possible</td>
<td>No</td>
</tr>
<tr>
<td>-Height</td>
<td>No</td>
<td>No</td>
<td>Potential indirect</td>
<td>Potential direct</td>
</tr>
<tr>
<td>Ice type</td>
<td>Potential</td>
<td>Yes with SMMR</td>
<td>Potential</td>
<td>Potential</td>
</tr>
<tr>
<td>Leads:</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>-Fractional area</td>
<td>Yes</td>
<td>No</td>
<td>Yes</td>
<td>Potential</td>
</tr>
<tr>
<td>-Orientation</td>
<td>Yes</td>
<td>No</td>
<td>Yes</td>
<td>Potential</td>
</tr>
<tr>
<td>Floe position</td>
<td>Yes</td>
<td>No</td>
<td>Yes</td>
<td>Potential</td>
</tr>
<tr>
<td>Surface melting</td>
<td>No</td>
<td>Yes</td>
<td>Potential</td>
<td>Potential</td>
</tr>
<tr>
<td>Surface temp.</td>
<td>Possible with infra-red</td>
<td>Indirectly</td>
<td>No</td>
<td>No</td>
</tr>
<tr>
<td>Ice thickness</td>
<td>No</td>
<td>Only from ice type</td>
<td>Only from ice type</td>
<td>Potential direct measurement of freeboard</td>
</tr>
</tbody>
</table>

Table 2.8 Capability of different satellite sensors to provide sea ice parameters specified in table 2.3.

It should be borne in mind that the capabilities of each instrument are limited by the coverage outlined in table 2.5. For example cloud cover and darkness seriously limit the ability of visible and infra-red observations to be used on a global scale. Although SAR can provide many measurements of value, difficulties in interpretation, and the limited amount of data currently available, make global studies difficult. Most studies of sea ice on a global scale have therefore depended on passive microwave data.

Satellite altimeters have the potential to improve on passive microwave measurements of two key parameters. Although the across track coverage of altimeters is poor, they have the potential to provide the location of the ice boundary, on a global scale, to an accuracy of a kilometre or so. In addition the great sensitivity of the altimeter to smooth surfaces may allow precise estimates of the percentage of open water and thin ice coverage at high ice concentrations (>90%), which are key to sea ice model inputs. Passive microwave
observations at high ice concentrations are limited by the uncertainties in ice emissivity.

If the measurements of surface roughness and elevation, which are provided by altimeters over the open ocean, can be extended into sea ice areas, then altimeters can potentially provide direct measurements of ice roughness and freeboard which are not available from any other sea ice sensor. Detailed information on ice ridging, leads and fractures are becoming more vital for sea ice model validation [Barry et al., 1984]. Ice thickness is a key variable in many sea ice models which are limited by a lack of suitable data [Zwally, 1984].

2.6 Summary

- Sea ice roughness and salinity vary according to the age of the ice cover.

- Sea ice cover shows significant spatial variation depending on the location of land masses and the actions of wind, ocean currents and waves.

- High resolution visible imagery can give detailed information on sea ice cover in small areas but wider coverage is limited by darkness and cloud cover.

- Infra-red observations provide wider coverage at lower resolution but are still limited to a great extent by cloud cover.

- Passive microwave data provide wide coverage and is unaffected by cloud cover but at the expense of resolution.

- Synthetic aperture radar has the potential to provide day and night all weather high resolution imagery but data interpretation is still a research subject. In addition technical problems will limit coverage by satellites in the near future.

- The analysis of radar altimeter data has been carried out by several different research groups but instrumental problems and very limited ground truth have proved limiting factors.
3.0 Introduction

In this chapter we will describe the operation of satellite borne altimeter systems. All space-borne altimeters flown so far were designed specifically for operation over the ocean and thus assumed a relatively flat surface with isotropic backscattering characteristics within the instrument footprint. This assumption causes serious problems over surfaces which do not generally satisfy these criteria, such as land and sea ice, and results in large errors in the on board estimates of geophysical parameters. To retrieve information from non-ocean surfaces it is necessary first to understand why problems occur. For this reason the operation of the on board processor is discussed in some detail.

We first outline the principles of altimetry and then concentrate specifically on the Seasat altimeter. Data discussed in later chapters originates either from the Seasat or Geosat altimeters which are essentially identical instruments. We discuss aspects of the Seasat altimeter on board processing and telemetry which are important to the study of altimeter returns over non-ocean surfaces. A more detailed description of the Seasat altimeter can be found in MacArthur [1978].

3.1 Principles of radar altimetry

3.1.1 Return pulse profile

The operation of the altimeter begins with the emission of a short pulse which expands as a spherical shell towards the surface. For a flat surface, once pulse reaches the point on the surface closest to the satellite, the area illuminated expands first to a circle and then to an annulus of constant area [Rapley et al., 1983].

For a flat, homogeneous surface the return waveform profile can be obtained from the convolution of three functions: the transmitted pulse profile, $S_r(t)$, the flat surface impulse response, $P_{fs}(t)$, and the probability density function for the height of specular points on the
surface, $q_s(t)$, [Brown, 1977]. The return waveform, $P_r(t)$, is obtained from:

$$P_r(t) = P_{fs}(t) \otimes q_s(t) \otimes S_r(t)$$  \hspace{1cm} 3.1

Where $\otimes$ indicates convolution. The height distribution function, $q_s(t)$, depends on the large (\sim m) scale roughness of the surface. Over the open ocean this can be extracted from the return waveform to provide information on metre scale roughness. The form of $P_{fs}(t)$ depends on the pointing of the satellite, the antenna pattern and the form of the backscattering coefficient $\sigma^0(\theta)$.

### 3.1.2 Modes of operation

Theoretically altimeters can operate in two distinct modes: beam limited operation or pulse limited operation. These two modes are discussed in more detail by Rapley et al. [1983]. In summary they state that, although beam limited operation is preferable where the radar backscatter and surface height varies on short spatial scales, all space borne altimeters so far flown have operated in pulse limited mode due to technical constraints on the size of antenna employed.

### 3.1.3 Waveform averaging

The radar return from a rough surface observed by the altimeter consists of returns from many small specular facets. The phase of the return from each facet will vary according to its exact range from the sensor. For a rough surface the returns from single facets will add up in a random manner resulting in 'fading noise' on the radar return. For a single return pulse the distribution of power in a single bin, $p_i$, will be exponentially distributed about the mean thus:

$$pdf(p_i) = \frac{1}{<p_i>} e^{-\frac{p_i}{<p_i>}}$$  \hspace{1cm} 3.2

where $pdf()$ indicates the probability density function of received power and $<p_i>$ is the expectation value (or mean for a large number of samples) of the power in sample $i$. 
To create a return waveform from which reasonable estimates of surface parameters can be extracted a large number of pulses are averaged. For the Seasat altimeter 50 pulses were averaged before processing by the on board tracking computer was carried out. Each telemetred waveform is the sum of two tracking loop updates and is therefore the sum of 100 individual return pulses. The variance in each waveform sample is therefore given by:

\[
\sigma^2[p_i] = \frac{\sigma^2[p_i]}{100} = \frac{<p_i>}{100}
\]

This variance will apply to all return echo samples which result from incoherent reflection.

3.2 The Seasat altimeter system

3.2.1 Full deramp processing

The Seasat altimeter uses full deramp processing which allows a long (3.2μs) transmitted signal to be compressed to an effective pulse length of 3.125ns. The advantage of this system is that it allows a high mean to peak power ratio whilst maintaining a short pulse length.

![Figure 3.1 Principle of Full Deramp pulse compression](image)

The principle of this method of pulse compression is shown in figure 3.1. The transmitted signal consists of a linear sweep over a 320 MHz frequency range centered about the
operating frequency of 13.5 GHz. The received signal is mixed with the transmitted signal to obtain an Intermediate Frequency (IF) signal. The power at a particular frequency in the IF signal corresponds to the power received at a particular delay time.

### 3.2.2 Seasat return waveform sampling

The intermediate frequency spectrum, produced by mixing the transmitted and received pulses, is sampled by a number of filters. This provides a series of samples of the return power in so called 'range bins'. Each range bin has a width equal to the transmitted pulse length, 3.125ns., corresponding to a range resolution of 0.478m. The main record consists of a series of 60 range bins, normally centered about the leading edge of the return, known as the range window. The convention adopted throughout this work is that the range gates shall be numbered from 1 to 60.

![Range window diagram](image)

**Figure 3.2** Seasat return waveform sampling. The delay time, off-nadir range and angle are shown for the case of a flat surface.

In addition to the 60 bins appearing in the range window a further 3 bins appear at the center of the range window to allow for higher precision tracking when the surface roughness is small. These bins are centred in the range window and correspond to bins at positions 29.5, 30.5 and 31.5. A further sample is made in a gate which samples the return power some 0.55° off-nadir. The primary purpose of this gate is to provide an estimate of off-pointing angle when the altimeter is operating over the open ocean. The sixty range bins and the attitude gate are shown schematically in Fig 3.2.
3.3 Pulse limited altimeter geometry

3.3.1 Range ring geometry

For a flat surface the power appearing in individual waveform bins originates from a series of annuli, centered on the nadir point, known as range rings. The range rings for the Seasat altimeter are shown in figure 3.3.

![Figure 3.3 Seasat range ring geometry](image)

Taking account of the Earth's curvature the outer radius of a range ring, $r_n$, which contributes power to bin $n$ is given by Rapley et al. [1983] using:

$$r_n = \sqrt{\frac{nc\tau h}{1 + \frac{h}{r_e}}}$$

where $c$ is the velocity of light, $h$ is the altimeter altitude, and $r_e$ is the radius of the Earth. The off-nadir angle for a flat surface, and for small angles is therefore given by:
3.5

The time delay $\tau'$ will increase with surface roughness as the return power is spread out. It value is given by:

$$\tau' = \sqrt{\tau^2 + \frac{16\sigma_h^2 \ln 2}{c^2}}$$  

3.2.2 Altimeter footprints

We now define three footprints of importance for satellite altimetry:

3.2.2.1 Pulse Limited Footprint

The Pulse Limited Footprint (PLF) defines the area which contributes to the leading edge of the radar return and is the limit of spatial resolution for measurement of surface height and roughness over a diffuse flat surface. Using equation 3.4, with $n=1$, $h=800$km, $r_e=6400$km and $\tau=3.125$ns we obtain $r_{PLF}=816$m.

3.2.2.2 Range Window Footprint

The range window footprint defines the area corresponding to power received within the range window. Again using equation 3.4, this time with $n=30$, we obtain $r_{RWF} = 4469$m. This footprint, for a flat surface and normal tracking, defines the area from which power appearing in the range window originates.

3.2.2.3 Beam Limited Footprint

For completeness we also define the Beam Limited Footprint (BLF) here which is taken to be the area over which the one way attenuation due to the antenna gain is less than 3dB. The nominal 3dB half width of the Seasat antenna pattern is $0.8^\circ$ giving a beam limited footprint
radius, $r_{BLF} = 11$km. It should be noted that the choice of a 3dB cutoff is arbitrary and does not preclude returns from outside this footprint from appearing in the range window.

3.4 Seasat altimeter on board processing

3.4.1 Seasat control loops

Variations in backscatter of around one order of magnitude normally occur over the ocean with a much larger variation, five or six orders of magnitude, possible over non-ocean surfaces (see chapter 6). Variations in the satellite altitude above the surface of the Earth can cause the range to the surface to vary over a range three orders of magnitude larger than that covered by the range window.

The processor on board the Seasat altimeter used two control loops to maintain reception of a signal from which useful information could be retrieved. The first, the AGC loop, was designed to adjust the gain of the receiver so that the signal level remained below saturation level whilst giving a good signal to noise ratio. The second, the height loop, was designed to maintain the leading edge of the return waveform, where most of the useful information appeared, in the centre of the range window. Both loops were updated at a rate of 20Hz. The loops are driven by the selection of various 'gates' which are made up from the summed power in one or more return waveform bins. The selection of these gates and the processing of the loops is discussed next.

3.4.2 Automatic Gain Control loop processing

The Automatic Gain Control (AGC) determines the gain applied to the signals recorded in the range gates. The value of the AGC can be adjusted from 0 to 63dB in steps of 1dB. The AGC loop is controlled by the AGC gate, which gives a mean count value in the waveform. The AGC gate value is computed from:

$$AGC_{GATE} = \frac{1}{33} \sum_{i=1}^{60} P_i$$

The loop is designed to force the AGC gate to an average count value of 46. The value of
AGC on the receiver is governed by a running average of the power in the AGC gate. The recursion relation used was:

\[ AGC_{n+1} = AGC_n + a \Delta AGC \]  \hspace{1cm} (3.8)

Where \( \Delta AGC \) is calculated from the difference between \( AGC_{\text{GATE}} \) and some reference value. For the normal track mode used the parameter \( a \) has a value of 1/16 providing a \(~0.8s\) integration time. The AGC thus also provides a smoothed measure of the return power observed and is used to derive the value of \( \sigma^0 \) over the ocean.

### 3.3.3 Height loop processing

The position of the range window is controlled essentially by a negative feedback loop. The loop attempts to minimise the height error calculated from a series of split gates centred around the leading edge of the range window. The operation of the height control loop is more complex than the operation of the AGC control loop. The first reason for this is that it must allow for a constant change in the altimeter range that commonly occurs over the ocean. The second is that optimum performance of the tracker is achieved by analysing waveform samples over the entire leading edge of the waveform, which varies with changing surface roughness.

The operation of the height loop begins with the selection of a so called 'gate triplet' - a series of three gates centred in the range window. These gates are made up by summing the power in various range bins. The width of these gates depends on the leading edge width estimated from an earlier return. Six different combinations are possible for the gate triplet and these are listed in table 3.1.
Table 3.1 Possible gate triplets selected by the Seasat onboard processor

<table>
<thead>
<tr>
<th>Triplet number</th>
<th>Gates Early (Ei)</th>
<th>Gates Middle (Mi)</th>
<th>Gates Late (Li)</th>
<th>Gate Width</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>30</td>
<td>30.5</td>
<td>31</td>
<td>1</td>
</tr>
<tr>
<td>2</td>
<td>29.5</td>
<td>30.5</td>
<td>31.5</td>
<td>1.5</td>
</tr>
<tr>
<td>3</td>
<td>28-29</td>
<td>30-31</td>
<td>32-33</td>
<td>2</td>
</tr>
<tr>
<td>4</td>
<td>25-28</td>
<td>29-32</td>
<td>33-36</td>
<td>4</td>
</tr>
<tr>
<td>5</td>
<td>19-26</td>
<td>27-34</td>
<td>35-42</td>
<td>8</td>
</tr>
<tr>
<td>6</td>
<td>7-22</td>
<td>23-38</td>
<td>39-54</td>
<td>16</td>
</tr>
</tbody>
</table>

Once the gate triplet is selected the processor computes the height error, Δh, using:

\[ Δh = AGC_{\text{GATE}} - M_i \]

where \( M_i \) (i=1 to 6) is the value of the middle gate which has been selected. Thus for an ocean return which is perfectly centred in the range window the mean power in the middle gate should be equal to the mean power in the entire range window and hence \( Δh \) is zero. This is illustrated in figure 3.4.

Figure 3.4 Gate triplet values for the case of an 'ocean' like return

The relationship between the offset of an ocean return waveform from the centre of the range window and \( Δh \) is shown by Rapley et al. [1983] and is fairly linear for an offset of around 5
bins. Outside this range the height error decreases for offsets to the left of the tracking point and levels out for offsets to the right.

The computed height error is used to produce a predicted height value for the subsequent return. The predicted height value, $h_{n+1}$, is found from the previous height estimate, $h_n$, using:

$$h_n = h_{n-1} + \alpha \Delta h_n + h_n \Delta t$$

where $\alpha$ is the tracker time constant, $\Delta h$ is the height error and $h'$ is the height rate and $\Delta t$ is the update interval of the loops. The height rate is calculated using a loop:

$$h_n = h_{n-1} + \frac{\beta \Delta h_n}{\Delta t}$$

both of these height loops are updated 20 times per second.

The height loop constants, $\alpha$ and $\beta$, can be adjusted to modify the performance of the tracker. Decreasing the value of these constants increases the agility of the tracker but also increases noise on the height signal. Increasing them reduces the overall noise but does not allow for more rapid variation.

Another function accomplished by the height tracking loop is to permit an on board estimate of the significant wave height. This is done by computing the ratio:

$$\frac{\frac{L_i - E_i}{L_i - E_6}}{L_6 - E_6}$$

and then comparing its value with a look up table.

In certain circumstances, where rapid changes in range or distorted waveforms occur, the leading edge of return waveform may be lost. This situation is known as 'Loss of Lock' and frequently occurs over land surfaces. To regain correct tracking the altimeter is put into an acquisition mode during which information regarding the surface is lost. Fortunately this
phenomenon rarely occurred when the Seasat altimeter observed sea ice although it seems to occur more frequently with data from Geosat. More detail on the mechanism of acquisition is given in Rapley et al. [1983].

The on board estimates of range and surface roughness thus depend critically on the correct operation of the altimeter tracking loops. As we shall see this cannot be assumed when the altimeter is operating outside areas of open ocean.

3.5 Processing of non-ocean like returns

3.5.1 Peaked return AGC loop processing

The AGC control loop, which provides a smooth measure of the backscattered power over ocean surfaces, suffers several problems when operating over sea ice surfaces. Figure 3.5 shows an example of a peaked return appearing in the Seasat range window. Evidently the AGC gate calculated from the mean power in all sixty bins does not correspond to the maximum peak backscatter value which is of importance in determining surface characteristics.

The AGC loop also acts to smooth the return power that is observed and therefore may not reflect the true variations in power that occur when the altimeter traverses areas where backscattered power is varying rapidly. When crossing from highly reflecting areas to less bright areas the count value in individual waveform bins will fall rapidly and the AGC will overestimate the return power. In these situations the signal to noise ratio will also decrease significantly due to the higher gain and consequently lower count values that appear in the range window. In the opposite situation, passing from a low to a high backscatter area the count values in the range window will rise dramatically causing much higher count values to appear. By analysing the return waveform the peak, unsmootheed, backscatter value can be extracted a problem addressed in chapter 5.

In some cases, where the bin count value reaches very high levels, saturation of the receiver can occur [Wingham and Rapley, 1987]. The digitisation of return samples is limited to 9 bits which are converted to count rates using the multiplying factor of 100/64. Thus for count values above around 800 (=2^9*100/64) individual filters will become saturated. This can result in truncation of the peak samples in a return waveform by up to 14dB.
3.5.2 Peaked return height loop processing

The problem of processing data which does not conform to the normal 'ocean' like return can be seen in figure 3.5.

![Diagram showing gate triplets for a peaked return](image)

**Figure 3.5** Gate triplets for the case of a 'peaked' return

The example shows a peaked return, typical of those observed over sea ice, with a leading edge near the centre of the range window with triplet 3 selected. The mean power appearing in the middle gate will be substantially greater than the value in the AGC gate which, considering equation 3.9, results in a large negative value of \( \Delta h \). This has the effect of 'kicking' the range window to the left (decreased range estimate) so that the subsequent return pulse will appear later in the range window. For the subsequent pulse the power appearing in the middle gate will be from the very low level noise that appears prior to the main return and much less than the AGC gate. This will cause a positive height error resulting in a slow restoration of the return leading edge to the centre of the range window. As soon as substantial power appears again in the middle gate the tracker will again kick and the sequence will be repeated. This phenomenon is known as 'tracker oscillation' and an example of the effect on range estimates is shown in figure 5.1. The deviation of the leading edge of the return waveform from the centre of the range window will also corrupt the estimates of significant wave height made by the altimeter.

3.5.3 Telemetry summing

Another aspect of the Seasat on board processing important to studies over non-ocean surfaces is the way the data were averaged before being telemetred. Although the height and
AGC loops were updated at the rate of 20 frames per second the waveform data were telemetred at the rate of 10 frames per second.

On board estimates of the parameters and the waveform profile are thus the sum of the results from two updates of the AGC and height loops. Although this presents no problem over the ocean, where telemetred values are varying slowly, it has more important consequences for rapid variations that occur over non-ocean surfaces. In particular, where two waveforms are summed during a rapid excursion of the range window, a double peaked waveform can result. Such a waveform is the sum of two waveforms created just as the tracker 'kicked'. The severity of this problem depends on the exact phasing of the tracking loops and telemetry, in some cases the return waveform is simply blurred. In addition, when the range window is moving rapidly, there may be a significant error in the predicted range to the centre of the range window (known as a 'height glitch', caused by telemetry summing [Rapley et al., 1987]. The approach adopted for processing data presented later (section 5.3) is to identify such occurrences and to delete the affected waveforms if necessary.

3.6 Geometric effects

In this discussion we discuss aspects of altimeter geometry which can affect the on board estimates of spacecraft range, backscattered power and surface roughness and also the return waveform shape.

3.6.1 Antenna mispointing

First we consider altimeter observations of a planar surface which is not normal to the direction of observation. This can be result of either instrument off-pointing or observations over areas of significant surface slopes. The former can occur over any type of surface including the ocean but the latter is restricted to observations over land surfaces. Over sea ice the importance of this effect will be the result mainly of off-pointing since elevations over sea ice surfaces generally follow the contours of the ocean surface. Data from Geosat suffers particularly from instrument off-pointing.

The expression for the flat surface impulse response , assuming a Gaussian form for the antenna polar diagram, where the instrument is mispointing or over a sloping surface is given
\[ P_{fs}(t) = F e^{\frac{4\pi ct \cos(2\zeta)}{\gamma h}} U(t) I_0 \left[ \frac{4\sqrt{\pi \sin^2 \zeta}}{\gamma h} \right] \]

where \( U(t) \) is the unit step function, \( I_0 \) is a modified Bessel function, \( \zeta \) is the angle between the antenna boresight and the normal to the surface and the constant \( F \) is given by:

\[ F = P_t \frac{G^2 \lambda^2 c \sigma_0 e^{-\sin^2 \zeta}}{4(4\pi)^2 h^3} \]

the value of \( \gamma \) depends on the half (3dB) width of the antenna beam, \( \theta_b \), and is given by:

\[ \gamma = \frac{4 \sin^2 \left( \frac{\theta_b}{2} \right)}{\ln(4)} \]

For \( \zeta = 0 \) the argument of the bessel function will be zero and the \( P_{fs} \) will be determined only by the antenna pattern. As the mispointing angle (or surface slope) increases the peak waveform power will decrease and the fall off, due to the antenna pattern, will become less steep. When the off-pointing angle reaches or exceeds approximately half the antenna beamwidth the return waveform will first become flat and will then have a positive slope with time. For sloping surfaces the range measurement will no longer correspond to the nadir point with important consequences for the range measurement [Partington, 1988]. This is not the case for simple instrument off-pointing and this is therefore not discussed further here.

3.6.2 Topographic variations

For surfaces with topography which varies considerably within the altimeter footprint more complex return waveforms can occur. Retrieval of height information requires reconstruction of the surface and methods for achieving this are the subject of current research [D.J. Wingham, personal communication]. Since topography of a significant scale is rare over sea ice this problem is not addressed in this thesis.
3.6.3 Backscatter variations

Variations in surface backscatter over the range window footprint also have important effects for the estimation of surface characteristics from altimeter data. The most critical of these is 'off-ranging' which occurs when a bright area appears away from the nadir point. This is particularly manifest when the altimeter passes from an area of high backscatter to a less bright area. Over sea ice this typically occurs at the transition from the main ice pack to fast ice but can also occur when the altimeter passes over thin ice or open water into rougher first or multi-year ice. Where the radar reflection from a small area dominates the return waveform the on board tracker will attempt to follow the bright target at it recedes from the centre of the range window. This is known as 'tracker snagging' and will result in an overestimate of the range to the nadir point for a short period. The resulting errors can be several metres and the identification of such occurrences is necessary for accurate retrieval of surface elevations over sea ice areas. In some instances this phenomenon can be used to advantage in mapping ice shelf boundaries [Zwally et al., 1987b].

3.7 The ERS-1 altimeter

The altimeter to be deployed on the ERS-1 satellite differs in several important aspects to that flown on board Seasat and Geosat. The operating frequency and compressed pulse length will be similar but waveform profiles will be telemetred at a rate of 20 Hz, thus overcoming the problems suffered due to the telemetry summing employed on Seasat and Geosat. Although tracking errors are still likely to occur over sea ice surfaces retracking of data will be considerably more reliable with the elimination of the height glitch phenomenon.

The ERS-1 altimeter range window width will also vary, being four times wider when in ice mode as compared to the normal ocean mode. This will provide samples of radar backscatter over wider incidence angles and may allow better discrimination for some types of sea ice cover (see discussion of MIZEX altimetry 2.4.4.5).
3.8 Summary

- Over a flat surface the return waveform profile from an altimeter depends on the transmitted pulse profile, the antenna pattern, the backscattering pattern and the surface roughness.

- Technical difficulties have resulted in the use of the pulse limited mode for satellite altimetry in preference to beam limited operation.

- The Seasat/Geosat altimeters use a technique known as pulse compression to achieve a high echo to signal noise ratio.

- The Seasat/Geosat altimeters provide a series of high resolution time gates of the received power from which much information, on the surface characteristics, can be extracted.

- The spatial resolution of the altimeter is not fixed and is related to surface characteristics such as backscattering pattern and surface roughness.

- The on board processor used on Seasat/Geosat assumed a reflection from an ocean surface.

- Reflections from non-ocean surfaces can severely compromise on board estimates of surface parameters provided by the altimeter.

- Retrieval of information can be achieved by reprocessing return waveforms but detailed knowledge of the processing loops and telemetry system must be considered.

- Varying surface topography and backscattering characteristics can seriously affect operation of the on board tracker. To retrieve precise surface elevation and backscatter measurements requires reprocessing of individual return waveforms.
Certain phenomena, such as instrument saturation and pulse blurring can prevent retrieval of accurate information. In such cases data should be edited from processing.
4.0 Introduction

In this chapter we consider how a radar altimeter can provide information about sea ice through the magnitude of the return power and the shape of the return waveform. The di-electric properties of sea ice are discussed briefly since observations point to the fact that surface roughness characteristics are primarily responsible for the variations observed.

The prediction of radar backscatter from natural surfaces relies on observed or theoretical surface roughness distributions. Over the open ocean, models for the surface roughness characteristics have been successfully used to predict radar backscatter at varying wind speeds (e.g. Barrick [1974]). No such models and very few observations exist which provide the detailed information on sea ice surface roughness needed by most models. Nevertheless a review of the basic principles and backscattering theories applicable to rough surfaces is needed, both to improve interpretation of data, and to determine which surface roughness parameters should be collected in field campaigns.

Model predictions of three waveform parameters are given; SIGPK - a measurements of peak backscatter, SIGTD - a measure of the fall off in power with incidence angle and SIGTR - a measure of the deviation of the return waveform from an exponential decay. The method for computing these parameters, which are used as the basis for studies presented in later chapters, is explained in section 5.8.

Airborne and ground measurements of the backscatter from sea ice concentrate mainly on off-nadir angles relevant to SLAR and SAR observations. Nadir observations are scarce although some do exist. Those which do show that reflections from calm water usually dominate near nadir. The situations under which water can be considered calm are also therefore discussed in detail.
4.1 Measurement of radar backscatter by a radar altimeter

A satellite borne radar altimeter, in addition to providing a measurement of the surface range, provides a measurement of the near normal incidence backscatter pattern. For a flat surface (see chapter 3) the power appearing in the range window, after the initial rise, corresponds to surface reflection at increasing angles away from nadir. For the ocean surface, which is essentially an isotropic scatterer over the narrow range of incidence angles sampled by the altimeter, the decrease in return power after the initial rise can be attributed mainly to the effects of the antenna pattern. Over most areas of sea ice, however, return waveforms show a sharp decrease in return power away from nadir.

The radar cross section of a target (RCS) is frequently used as a direct measure of the return signal that will be seen by a radar. For a monostatic radar (i.e. one where the same antenna is used to transmit and received the radar pulse) the radar equation is given by:

\[ P_r = P_t \frac{G^2 \lambda^2 L_{\text{atm}}}{(4\pi)^3 h^4} \text{RCS} \]

where \( P_t \) is the transmitted power, \( P_r \) is the received power, \( G \) is the antenna gain, \( \lambda \) is the radar wavelength, \( L_{\text{atm}} \) is the loss due to atmospheric attenuation and \( h \) is the range of the target. The radar backscatter co-efficient, \( \sigma^0 \), of a surface is then defined as the average radar cross section per unit area of the surface. For a given part of the surface of area \( A \) and radar cross section RCS, \( \sigma^0 \) is given by:

\[ \sigma^0 = \frac{\text{RCS}}{A} \]

For a pulse limited radar altimeter we are interested in relating the backscatter co-efficient to the return power appearing in a waveform sample. For a flat surface the power appearing in any waveform sample will come from a range ring with an area equal to the pulse limited footprint (PLF). The radius of the PLF is given by Rapley et al. [1983]:

\[ r_{\text{PLF}} = \sqrt{c h t} \]
the area $A$ of the surface over which the power received appears in a single range ring is thus:

$$A = r^2_{PLP} = \pi c h$$

4.4

substituting equations 4.2 and 4.4 in equation 4.1 the power appearing in a single waveform sampler, $P_r$, is given by:

$$P_r = P_t \frac{G^2 \lambda L_{atm} c \tau}{4(4\pi)^2 \frac{h^3}{3}} \sigma^0$$

4.5

This is the link equation for a pulse limited altimeter.

4.2 Di-electric properties of sea ice

The principles governing the reflection of electromagnetic waves are widely covered in Hecht and Zajac [1973]. Although these are well known by those with a physics background some aspects are worth highlighting since they are the basis of the backscattering theories described later and allow some understanding of altimeter data where theories may not be available.

The interaction of an electromagnetic wave with the media through which it is travelling is determined by two constants; the relative permittivity, $\varepsilon_r$, related to electrical properties and the relative permeability, $\mu_r$, related to magnetic properties. For most substances, with the exception of ferro-magnetic materials, the value of $\mu_r$ is 1. The Fresnel power reflection coefficient, $R^2$, of an electromagnetic wave incident on a plane boundary is, therefore, determined solely by the di-electric constant and is given by:

$$|R(0)|^2 = \left[ \frac{1-\sqrt{\varepsilon_r}}{1+\sqrt{\varepsilon_r}} \right]^2$$

4.6
At a plane boundary the amount of energy reflected by a surface is determined by the value \( R^2 \) with the remainder being transmitted into the medium. Transmitted radiation will undergo further scattering within the medium until finally being absorbed.

The di-electric properties of ice and snow, including sea ice, are reviewed by Drinkwater [1987]. Measurements of the di-electric properties of ice can also be used to calculate the Fresnel reflection co-efficient using equation 4.6. For pure ice the reflection co-efficient is very constant over the entire microwave region with a value around -11dB. Sea ice is, typically, a more complex medium because of the inclusion of brine pockets which act to increase the overall reflection co-efficient. Drinkwater quotes an equation for the relative permittivity of sea ice, \( \varepsilon'_{si} \) in terms of the relative brine volume \( V_b \):

\[
\varepsilon'_{si} = \frac{\varepsilon_i}{1 - 3V_b}
\]

where \( \varepsilon_i' \) is the relative permittivity of pure ice. The relative volume of brine is obtained by combining the brine salinity with the volume of brine enclosed. Brine salinity will depend on the age of the ice with older ice having a lower salinity than younger ice due to drainage (see section 2.1). The volume of brine enclosed will also generally increase with both temperature and density. Although exact values cannot be assumed approximate values can be obtained using empirical formulas. Table 4.1 gives typical values from Drinkwater for the resultant reflection coefficients for first year ice of density 850 kgm\(^{-3}\) and multi-year ice of density 750kgm\(^{-3}\). The ranges of salinity assumed are 4-16 parts per thousand for the first year ice and 0.6-1 parts per thousand for the multi-year ice.

<table>
<thead>
<tr>
<th>Temperature (°C)</th>
<th>First year ice</th>
<th>Multi-year ice</th>
</tr>
</thead>
<tbody>
<tr>
<td>-30</td>
<td>-11.2 to -10.8dB</td>
<td>-12.1dB</td>
</tr>
<tr>
<td>-20</td>
<td>-10.8 to -8.4dB</td>
<td>-12.1 to -11.9dB</td>
</tr>
<tr>
<td>-10</td>
<td>-10.4 to -7.6dB</td>
<td>-12.0 to -11.7dB</td>
</tr>
<tr>
<td>-2</td>
<td>-7 to -4dB</td>
<td>-11.2 to -10.2 dB</td>
</tr>
</tbody>
</table>

Table 4.1. Fresnel reflection coefficients for First and Multi-year ice with varying temperatures [Drinkwater, 1987].
For the more saline first year ice the dependence on temperature is more critical. Density variations have a lesser effect, generally increasing the reflection co-efficient by 1 - 2 dB between 650 - 850 kgm$^{-3}$. For both types of ice the reflection co-efficient increases most rapidly near freezing tending to the, virtually constant, -2dB reflection co-efficient of sea water at freezing.

The reflection coefficient governs the fraction of reflected energy from an electromagnetic wave incident on a plane surface. These results show that the reflection co-efficient can vary by up to 10dB depending on ice and surface temperature. Normal incidence backscattering values observed by the Seasat altimeter vary over a much larger dynamic range (~60dB) as shown in chapter 6. Di-electric properties are therefore a small, although not negligible, factor in determining the radar return observed by a satellite borne radar altimeter. To understand the much larger variations that are observed requires consideration of the possibility of coherent or partially coherent surface reflections.

### 4.3 Reflection from a perfectly flat surface

Over a perfectly flat surface the phase of the radiation reflected from different parts of the surface will vary according to distance from the nadir, or sub-satellite point. At some distance away from the nadir point the radiation will have undergone a phase change of $\pi$ radians compared with the first return. A circle with a radius equal to this distance defines the first Fresnel zone. The difference in range between the nearest point to the altimeter and the outer radius, $R_{FZ}$, of the first Fresnel zone is $\lambda/4$. This can be obtained from altimeter geometry using:

$$
R_{FZ} = \sqrt{\frac{\lambda h}{2 \left[ 1 + \frac{h}{a} \right]}} 
$$

Where $a$ is the Earth's radius. For the Seasat altimeter with $h = 800km$ and $\lambda = 2.2cm$ the radius of the first Fresnel zone, $R_{FZ}$, is $88m$. For an airborne platform with $h = 10km$ and $\lambda = 2.2 cm$ the radius of the first Fresnel zone, $R_{FZ}$, is 10.5m.
Further zones are defined for each change in sign of the reflected electromagnetic wave corresponding to a phase change of $\pi$. These are shown in figure 4.1. The power returned to a radar altimeter will originate from the first Fresnel zone only.

![Figure 4.1](image)

**Figure 4.1** Fresnel zones for the Seasat altimeter.

Summing over an infinite number of Fresnel zones the total electromagnetic field at the point of observation is equal to the sum of the contributions from each of the Fresnel zones:

$$E = |E_1| - |E_2| + |E_3| - \ldots |E_m|$$

4.9a

We can see that for an infinite number of zones a problem arises if the power from each zone is the same as the result depends on whether $m$ is chosen to be odd or even. In practice the power from each zone decreases due to the Kirchoff obliquity factor which goes to zero as $m$ goes to infinity. Hecht and Zejac [1973] show that the resultant field, $E$, is given by:

$$E \sim \frac{|E_1|}{2}$$

4.9b

The return power seen at the altimeter, therefore, originates from the first Fresnel zone only.
for a perfectly flat surface. The received power $P_r$ for a radar altimeter over such a surface is given by:

$$P_r = P_t \frac{G^2 \lambda^2}{64 \pi^2 h^2} |R(0)|^2$$

4.10

Comparing with equation 4.5 we find that the effective backscattering coefficient from such a surface is (Also given by Ulander [1987b]):

$$\sigma^0 = \frac{h}{c \tau} |R(0)|^2$$

4.11

taking values from the Seasat altimeter of $h = 800\text{km}$ and $\tau = 3.125\text{ns}$ and using $|R(0)|^2 = -2\text{dB}$ we obtain $\sigma^0 = 57.3\text{ dB}$. For an airborne system with $h = 10\text{km}$, the corresponding value is $38.3\text{ dB}$.

4.4 Reflection from a collection of facets

4.4.1 Diffraction theory

We now consider methods used to compute the electromagnetic field resulting from an individual facet on the surface. Diffraction theory is widely employed in optics problems to compute the radiation observed when coherent radiation is incident on a screen containing one or more apertures. Figure 4.2 shows how the radiation occurring at different points in a plane of observation can be calculated.

The disturbance, $dE$, arriving at point $P$ from a differential area $dS$ is given by:

$$dE = \left[ \frac{\varepsilon_A}{x} \right] e^{i(\omega t - kx)} dS$$

4.12

where $\varepsilon_A$ is the source strength per unit area and $x$ is the distance from $P$ to $dE$. This
expression can be integrated to compute the total electrical field at any point. For regular apertures analytical solutions are possible.

![Diagram of electrical field at a point from an extended source](image)

**Figure 4.2** Electrical field at a point from an extended source

For a circular aperture the power observed at a point P which lies at an angle \( \theta \) from the axis of the aperture is given by:

\[
I(\theta) = I(0) \left[ \frac{2J_1(k\sin\theta)}{k\sin\theta} \right]^2
\]

4.13

Where \( J_1 \) is a Bessel function of the first kind (1st order) and \( I(0) \) is the power at a distance \( R \) on the axis of the aperture given by:

\[
I(0) = \frac{\varepsilon_A^2 A^2}{2R^2}
\]

4.14

where \( A \) is the area of the aperture.

From equation 4.14 we can immediately see the dramatic effects that the presence of large coherent areas can have on the power detected by the altimeter. A flat area with a radius of 10m will return \( 10^4 \) times as much power as a flat area with a radius of 1m.

Fraunhofer diffraction can only be used with observations made in the 'far field'. The point of observation is said to be in the 'far field' if the phase of the radiation varies by less than about \( \pi/2 \) over the coherent areas. We can express this mathematically by:
Where \( R \) is the distance from the point of observation to the aperture, \( \lambda \) is the wavelength and \( r \) is the radius of the aperture.

For space borne altimeters with \( R = 800 \text{km} \) and \( \lambda = 2.2 \text{ cm} \), the far field assumption can be used with surface facets with a linear dimension \( r << 130 \text{ m} \). For an airborne altimeter, observing at a similar wavelength, but at a reduced range of 10 km the requirement becomes \( r << 15 \text{ m} \). Thus in situations where sea ice results in coherent returns over areas larger than a few metres it may not therefore be possible to directly compare airborne and satellite radar altimeter observations.

### 4.4.2 Phase coherency between different facets

If we wish to consider the reflected signal from a surface other than a plane boundary, phase interference of the radiation reflected from different parts of the surface becomes important. Most theories assume that the surface is made up of a series of facets, a facet being a small area on the surface from which the radiation received at the point of observation is in phase. For a collection of \( N \) facets, each resulting in an electrical field, of strength \( E \) and phase \( \phi \), at the receiver, Hecht and Zejac [1973] give the resultant electromagnetic field as:

\[
\begin{aligned}
E_0^2 &= \sum_{i=1}^{N} E_{0i}^2 + \sum_{i=1}^{N} \sum_{j>i}^{N} E_{0i} E_{0j} \cos(\phi_i - \phi_j) \\
&= NE_{01}^2
\end{aligned}
\]

If the phases from all the facets are randomly distributed then the second term in equation 4.16 will disappear and the sum, assuming all the waves have the same amplitude, will be:

\[
E_0^2 = NE_{01}^2
\]

This is known as incoherent reflection.
Now consider the situation when all the reflections are in phase, the power detected will then be the square of the sum of all the reflections:

\[ E_0^2 = \left( \sum_{i=1}^{N} E_{0i} \right)^2 = N^2 E_{01}^2 \]  

This is known as coherent or 'specular' reflection. Since the number of facets involved in reflection from a surface is generally very large the magnitude of the coherent return will be much larger than that for the incoherent return.

In addition to the above two situations we can also have reflections that are partially coherent or 'quasi-specular'. Consider a situation where we have a surface consisting of \( M \) areas each of which contains \( N \) facets. If the return from each facet in an area is in phase but the returns from separate areas are out of phase than the return power will be:

\[ E_0^2 = MN^2 E_{01}^2 \]  

As the size of individual areas increases the number of facets in each area, \( N \), will increase and thus the total received power will increase.

Although the above discussion may be considered simplistic, given the complexity of natural surfaces, it is a key principle to be applied to the problem of normal incidence backscatter over sea ice.

4.4.3 A facet model for normal incidence backscatter from first-year sea ice

Brown [1982] has developed a backscatter model based on a number of equiheight areas randomly distributed within the footprint of a space borne altimeter. All the areas on the surface will generate Fraunhoffer diffraction patterns dependent on their individual shapes and sizes. If the areas have a random transverse location then the phase of the returns from the areas will also be randomly distributed. The resultant radiation from each of the areas will add incoherently at the altimeter and the diffraction pattern obtained will simply be the sum of the
patterns from all the apertures. The fall off in the backscatter away from normal incidence will therefore be determined by the shape and size of a single aperture.

The exception to this is for areas lying at normal incidence where the radiation will add coherently at the altimeter. The Brown model therefore has two components corresponding to the coherent and incoherent reflections from the small flat areas:

\[
\sigma^0_{	ext{coh}} = |R(0)|^2 k^2 4\pi \langle \eta \rangle^2 \langle A^2 \rangle
\]

\[
\sigma^0_{	ext{inc}} = \frac{R(0)^2 k^2 \langle \eta \rangle \langle A_n^2 f_n^2 \rangle}{\pi}
\]

Where \(\langle \rangle\) denotes the expectation value, \(\langle \eta \rangle\) is the number of facets per unit area and \(A_n\) is the area and \(f_n\) the diffraction patterns of individual facets.

The incoherent component depends on the shape and size of the individual facets. Brown assumes that the areas responsible for the reflection are circular with radii that are Rayleigh distributed with the result that:

\[
\langle A_n^2 f_n \rangle = 2\pi^2 \sigma_p^4 \left[ I_0(\mu) - I_1(\mu) \right] e^{-\mu}
\]

where \(\mu = 2k^2 \sigma_p^2 \sin^2 \theta\), and \(\sigma_p^2\) is the second moment of the radius distribution.

Figure 4.3 shows the value of waveform parameters against \(\sigma_p^2\), a constant area coverage of about 10\% is assumed by setting \(\langle \eta \rangle = 0.1/\pi/\sigma_p^2\). Brown compares his model to observations made by the GEOS-3 altimeter, to extract surface parameters \(|R(0)|^2\), \(\langle \eta \rangle\), and \(\langle \eta \rangle A_n\) over sea ice. No surface measurements to validate these observations are presented and so the physical validity of the parameters extracted using this model must be viewed with some caution.
4.5 Rough surface scattering

At a plane boundary the reflected wave is a function of the electrical properties of the incident and reflective materials alone. In general, however, the radar reflection from natural surfaces cannot be computed from knowledge of the electrical properties alone due to superimposed roughness.

4.5.1 The Rayleigh criterion

The first problem to be addressed is how to decide whether a surface can be considered rough or smooth. Rayleigh proposed that a surface could be considered smooth if the phase difference, $\Delta \phi$, between radiation reflected from the extremes of surface roughness was less than $\pi/2$. This concept is illustrated in figure 4.4.

If we consider two rays incident on different parts of the surface separated in height by $\Delta h$ at an identical incidence angle $\theta$. The path difference between the two rays $\Delta R$ is then given by:
\[ \Delta R = 2\Delta h \sin \theta \]  \hspace{1cm} (4.22)

the phase difference between the two rays arriving at the receiver will therefore be:

\[ \Delta \phi = \frac{4\pi \Delta h \cos \theta}{\lambda} \]  \hspace{1cm} (4.23)

Where \( \lambda \) is the wavelength. If we require that \( \Delta \phi < \pi/2 \) then \( \Delta h \) must satisfy:

\[ \Delta h < \frac{\lambda}{8\cos \theta} \]  \hspace{1cm} (4.24)

Figure 4.4  The Rayleigh criterion for smooth surfaces. The surface is considered smooth if the path difference between rays striking extreme points on the surface is less than \( \lambda/8 \), where \( \lambda \) is the wavelength of observation.

For a space borne altimeter operating at 13.5 GHz at normal incidence this implies that for a surface to appear smooth requires a \( \Delta h < 3 \text{mm} \). A more stringent test is the Fraunhoffer criterion requires that \( \Delta \phi \) should be less than \( \pi/8 \). The value of \( \Delta h \) must then satisfy:

\[ \Delta h < \frac{\lambda}{32\cos \theta} \]  \hspace{1cm} (4.25)

Again using values of \( \theta \) and \( \lambda \) for a space borne altimeter we require a \( \Delta h \) of less than 0.75mm. Although the Rayleigh and Fraunhoffer criteria provide some quantitative indication of whether a surface can be considered smooth, in reality there is no exact dividing line between rough and smooth surface. What these results do indicate is that the critical roughness scale for altimeter observations lies in the mm region.
4.5.2 Small perturbation theory

When an incident wave impinges on a surface with a height deviation less than the Rayleigh criterion constructive interference of the reflected waves will occur. For small scale (i.e. still satisfying the Rayleigh criterion) roughness superimposed on such a surface this effective backscatter is modified to:

\[ \sigma^0 = \frac{h}{c\tau} |R(0)|^2 e^{-K_0} \]

where \( K_0 = 4k^2\sigma_h^2 \), \( k \) is the radar wavenumber and \( \sigma_h \) is the standard deviation of surface height.

![Graph showing the relationship between coherent backscatter and surface roughness.](image)

**Figure 4.5** Plot of peak backscatter versus standard deviation of surface height, \( \sigma_h \), for \( h = 800\text{km}, \tau = 3.125\text{ns} \) and \( |R(0)|^2 = 1 \) (from equation 4.26).

Figure 4.5 shows the dramatic fall of coherent power with increasing surface roughness. This graph emphasises the importance of mm scale surface roughness in determining the reflection of electromagnetic radiation with a wavelength of a few cm.
4.5.3 Physical optics approximation

For surfaces which do not satisfy the Rayleigh criterion rough surface scattering must be used. As yet no general and exact theory to describe radar backscatter from rough surfaces has been developed. All theories so far developed require assumptions to be made about the nature of the surface to make the problem tractable. The most common formulation used in the calculation of scattering from rough surfaces is the Kirchoff or physical optics approximation. The basic assumptions of this method [Beckmann and Spizzichino, 1963] are:

(i) The radius of curvature of the scattering elements is taken to be much greater than the wavelength of the incident radiation

(ii) Shadowing effects are neglected

(iii) Only the far field is calculated

(iv) Multiple scattering is neglected

Assumption (i) can also be expressed mathematically as:

\[
kl > 6.0 \\
I^2 > 2.76 \sigma_h^2 \lambda
\]

where \( k \) is the wavenumber (\( k=2\pi/\lambda \)), \( I \) is the correlation length of the surface and \( \sigma_h \) is the standard deviation of surface height. Two different types of approximation are used by Ulaby et al. [1981] to obtain analytic solutions to the Kirchoff scattering equations:

(i) Stationary phase approximation - where the standard deviation of surface heights is large compared with the wavelength (\( k\sigma_h >2.0 \)) - i.e. there is no coherent component. With this approximation we obtain:

\[
\sigma^0(\theta) = \frac{|R(0)|^2}{2s \cos \theta} e^{-\left(\frac{\tan^2 \theta}{2s^2}\right)}
\]

where \( s \) is the Root Mean Square (RMS) surface slope.
Figure 4.6 shows the values of SIGPK, SIGTD and SIGTR predicted for different RMS surface slopes using equation 4.28. It can be seen that the waveform shape changes from an 'ocean like' return only where the RMS surface slope falls below -0.01 (~0.5°). At this point the predicted peak backscatter is well above 30dB, assuming a return from 100% of the surface. This peak backscatter level would be less for only partial coverage or if new ice, with a lower reflection coefficient, were being observed. As can be proved theoretically the fall of in return power is very close to exponential, resulting in a value for SIGTR of near zero for all cases.

\[ \sigma_0(\theta) = \frac{|R(0)|^2 \sec^4 \theta}{s^2} e^{-\left( \frac{s \tan \theta}{s} \right)^2} \]

Equation 4.28 gives the expression for a surface with a Gaussian surface slope density. Brown [1979] has suggested that for low sea state conditions an exponential model for the surface slope density may be more appropriate. The expression for backscatter assuming such a model is given by Barrick [1968] :

Figure 4.7 shows the values of SIGPK, SIGTD and SIGTR predicted for different RMS surface slopes using equation 4.29. Values for the peak backscatter are somewhat larger for the same surface slope. Deviation of the return waveform shape from 'ocean like' also occurs somewhat
earlier although the rise in the value of SIGTD is less rapid. A significant difference is observed in the predicted values of SIGTR, which now show small positive values for small surface slopes.

Figure 4.7 Waveform parameter predictions for the Specular point theory stationary phase approximation with Gaussian distribution of surface heights, from equation 4.29. ($|R(0)|^2 = -2dB$)

(ii) Scalar approximation - for surfaces with small slopes ($s<0.25$) and a medium or small standard deviation in surface heights. This model assumes a Gaussian auto-correlation function for the surface. With this approximation we obtain:

\[
\sigma^0_{ch} = \frac{h |R(0)|^2}{c \tau} e^{-K_0} \\
\sigma^0_{re} = |R(0)|^2 (k \cos \theta)^2 e^{-K_0} \sum_{n=1}^{\infty} \frac{K_0^n}{n!} e^{-\left[\frac{(k \sin \theta)^2}{n}\right]} 
\]

where $K_0 = 4k^2 \sigma_h^2$ near normal incidence and $l$ is the correlation length.

Figure 4.8 shows the predicted value of each of the waveform parameters on a separate graph. This time values are given, both for different standard deviations of surface height, and for different correlation lengths. The peak backscatter value is observed to deviate from a linear increase as the standard deviation of surface height falls below 6mm and the coherent component starts to become important.
Figure 4.8a Specular Point theory (Scalar approximation) predictions for SIGPK.

Figure 4.8b Specular Point theory (Scalar approximation) predictions for SIGTD.

Figure 4.8c Specular Point theory (Scalar approximation) predictions for SIGTR.

Figure 4.8 Waveform parameter predictions for Specular point theory using the physical optic model, scalar approximation.
Both SIGTD and SIGTR parameters show values above zero only where the standard deviation of surface height is less than 6mm and where the correlation length is short. At longer correlation lengths the quasi-specular component starts to become more important resulting in changes in the waveform shape above 6mm surface roughness (i.e. entering the regime where the stationary phase approximation becomes valid).

4.5.3 Two scale models

Two scale models combine small perturbation and physical optics theories. The ocean surface is divided into a large scale, gently undulating height fluctuation \( h_L \) and a small scale surface height fluctuation \( h_s \). The wave height spectrum, \( \Psi \), for the total surface is given by the sum of the spectra of the long \( \Psi_L \) and short \( \Psi_s \) wavelength components:

\[
\Psi(k_x, k_y) = \Psi_L(k_x, k_y) + \Psi_s(k_x, k_y)
\]

where the two spectra are divided by a transition wavenumber, \( k_d \), such that:

\[
\Psi_L(k_x, k_y) = \Psi(k_x, k_y) \quad \text{for } k < k_d
\]
\[
\Psi_s(k_x, k_y) = \Psi(k_x, k_y) \quad \text{for } k > k_d
\]

The backscattering co-efficient can be calculated by combining the backscatter from the long wavelength part of the spectrum \( (\sigma^0_{QS}) \), derived using the physical optics approximation theories described previously, with the backscatter due to the short scale variation, derived from Bragg scattering. Durden and Vesecky [1985] give the resultant backscattering co-efficient as:

\[
\sigma^0(\theta) = e^{-K_0} \sigma_{QS}(\theta) + \int \frac{\sigma_{Bragg}(\theta, \psi, \delta)}{\cos \psi \cos \delta} p(\tan \psi, \tan \delta) d(\tan \psi) d(\tan \delta)
\]

The first term in this equation is obtained from the product of \( K_0 = 4k^2 \sigma_h^2 \) (as defined previously) and the equations for \( \sigma_{QS} \). The second term is derived by integrating the Bragg
scattering component ($\sigma_{\text{Bragg}}$) and slope density function ($p(\tan \psi, \tan \delta)$) over all angles of tilt defined by $\delta$ in the x direction and $\psi$ in the y direction.

At angles close to normal incidence the second term in equation 4.33 will be negligible and therefore the observed effect will be a decrease in the backscatter predicted by Quasi-specular scattering as small scale roughness increases. This theory is included to demonstrate that small scale roughness may act to decrease the overall backscatter near normal incidence in a manner which may not be detected in altimeter observations. This may seriously hamper attempts to quantitatively relate observations of the magnitude of radar backscatter in altimeter data to some geophysical parameters such as ice concentration (e.g. Drinkwater [1987]).

### 4.6 Empirical models and observations of radar backscatter from sea ice

Ulaby et al. [1986] reviewed the theory and observations of radar backscatter from sea ice. They admit that the theory of radar backscatter from sea ice is not well developed but quote a simple model by Kim [1984] which gives the radar backscatter from sea ice as:

$$
\sigma^0(\theta) = \sigma^0_s(\theta) + \Gamma^2 \sigma^0_v(\theta)
$$

where $\sigma^0_s$ is the backscatter due to the surface, $\Gamma$ is the transmission co-efficient and $\sigma^0_v$ is the volume scattering component. Kim shows that for First Year sea ice the backscatter from volume scattering is around -20dB but increases for multi-year ice, approaching 0dB in some cases.

Only a few observations of radar backscatter of sea ice at normal incidence exist. Figure 4.9 shows measurements at 13.3 GHz off the coast of Point Barrow, Alaska, during March 1970 Parashar et al. [1974]. Ice was divided into seven different categories, according its thickness which was estimated using aerial photography. The thickness and salinity of the different categories is given in table 4.2.

Given these results, in cases where the return echo of a 13GHz altimeter is dominated by reflections from the surface of ice floes, we would expect the peak backscatter value to be
lower than that normally observed over the ocean. In fact, as noted in section 2.3.6, radar altimeter observations of sea ice normally show much higher backscatter values (30-50dB in some areas) than are normally seen over the ocean. We are therefore led to conclude that the majority of radar altimeter observations of sea ice are dominated, either by areas of open water, or very new ice which is smoother than that observed in this experiment. In later chapters it will be demonstrated that backscatter values comparable to those seen in figure 4.6 are observed only where 100% ice cover exists within the altimeter footprint.

![Figure 4.9](image)

**Figure 4.9** Measurements of radar backscatter from sea ice at 13.3 GHz of point Barrow, Alaska during March 1970 [Parashar et al., 1974].
<table>
<thead>
<tr>
<th>Category</th>
<th>Ice type</th>
<th>Thickness (cm)</th>
<th>Surface salinity (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Open water</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>New ice</td>
<td>0-5</td>
<td>19</td>
</tr>
<tr>
<td>3</td>
<td>Thin young ice</td>
<td>5-18</td>
<td>13</td>
</tr>
<tr>
<td>4</td>
<td>Thick young ice</td>
<td>18-30</td>
<td>10.8</td>
</tr>
<tr>
<td>5</td>
<td>Thin First-year ice</td>
<td>30-90</td>
<td>7.42</td>
</tr>
<tr>
<td>6</td>
<td>Thick First-year ice</td>
<td>90-180</td>
<td>5</td>
</tr>
<tr>
<td>7</td>
<td>Multi-year ice</td>
<td>180-360</td>
<td>1</td>
</tr>
</tbody>
</table>

**Table 4.2** Thickness and salinity of different ice categories shown in figure 4.9.

### 4.7 Surface roughness in sea ice areas

Surface observations of the radar backscatter, shown in figure 4.9 and table 4.2, exhibit backscatter levels much lower than are normally observed in satellite altimeter data over sea ice. To endeavor to explain this observation we now consider the surface roughness of ice and water surfaces in sea ice zones.

![Roughness components of a sea ice covered ocean](image)

**Figure 4.10** Roughness components of a sea ice covered ocean.

Figure 4.10 shows the different components which will contribute to the overall roughness of an ice covered ocean. The water surface roughness is divided into both large and small scale roughness in keeping with the two scale model mentioned in section 4.5.3. From a physical point of view the large scale component will result from long wavelength swell that propagates into the pack and the small scale component represents the small ripples generated by local action of the wind.
It should be noted that not all the components shown in figure 4.10 will be present under all circumstances. For example meltponds will only be present where ice exists that has undergone a period of melting, and the long wavelength ocean roughness may be insignificant far into the ice pack due to the attenuating effect of ice floes.

4.7.1 Surface roughness of the ocean in sea ice areas

The effects of sea ice cover on ocean surface roughness have been the subject of both observational and theoretical work. The presence of sea ice will modify the wind generated wave spectrum in two distinct ways. Long period swell is damped out by ice floes although longer wavelengths may penetrate many kilometres into the pack. Short wavelength components of ocean surface roughness will be completely damped by the presence of ice floes but wind action within the ice pack and ice floe collisions may allow some regeneration at high frequencies [Wadhams et al., 1986].

4.7.1.1 Swell attenuation by sea ice

It is well established that long period swell incident on a pack ice edge will be subject to damping by the presence of ice floes both through observations and theoretical work [Wadhams, 1973; Wadhams, 1975; Wadhams, 1978; Squire and Moore, 1980, Wadhams et al., 1986]. Wadhams [1978] calculates the energy density for a given frequency, $G_x(f)$ at a distance $x$ into the pack using:

$$G_x(\omega) = G_0(\omega) e^{-2\alpha_x x} \quad 4.35$$

where $\alpha_x$ is the amplitude attenuation co-efficient and $G_0$ is the initial energy density. Values for $\alpha_x$ were obtained by observation and normally lies in the range $0.4 - 1.1 \times 10^{-4} \text{ m}^{-1}$ increasing with frequency. The highest values in this range were observed for very compact ice regimes where the rapid energy loss was attributed to the greater number of ice floe collisions.
Figure 4.11 Decrease in significant wave height and peak wave spectral frequency with distance into the ice pack (from tabulated values given in Wadhams et al. [1986]).

Figure 4.11 shows a graph of significant wave height (SWH), and peak wave spectrum frequency observed at different distances into the ice pack. The wave height is seen to decay rapidly to a few mm. If small scale surface roughness were absent this would imply coherent reflection from water areas within the pack. The peak wave spectrum frequency is also observed to fall with increasing distance into the pack demonstrating that longer wavelength components will penetrate further into the ice pack.

4.7.1.2 Wave generation by wind in the ice pack

In the interior of the ice pack it is predicted that the effects of swell will be negligible. Small scale roughness is completely damped out in a very short distance but may be regenerated by wind action in areas of open water. Given the fact that the previous discussion has pointed to areas of smooth water as a highly probable source for the high power peaked waveforms observed in altimeter data the mechanism of wind wave regeneration may be critically important. Wadhams [1983b] calculates the short wavelength roughness generated within the ice pack to predict the generation of ice bands by the selective action of wave generation by the wind. This calculation, although approximate, can be used to provide some estimate of the conditions under which areas of open water within the ice pack will produce coherent reflections.
Wadhams estimates the RMS surface roughness, $\sigma_h$, using the equation:

$$\sigma_h = 0.01265 U_* \sqrt{\frac{X}{g}}$$

4.36

Where $U_*$ is the friction velocity, $X$ is the fetch and $g$ is the acceleration due to gravity. Taking $g=9.8 \text{ms}^{-2}$ and assuming the wind speed $U_a = 20 U_*$, the standard deviation of surface height versus fetch for various wind speeds, calculated from equation 4.37, is shown in figure 4.12.

The results in figure 4.12 show the importance of wind speed in determining whether an area of open water of a given width will appear coherent to the altimeter. For a frequency of 13GHz this requires that $\xi_{\text{rms}} \leq 2.88 \text{mm}$ and $\xi_{\text{rms}} \leq 0.72 \text{mm}$ for the Rayleigh and Fraunhoffer criteria respectively. Although these relationships are only approximate they at least give a first order indication of circumstances under which open water within the ice pack is likely to produce a near specular return.
4.7.2 Surface roughness of grease ice

A field and laboratory study of wave damping by grease ice was carried out by Martin and Kaufmann [1981]. Laboratory results show that grease ice of 100mm thickness can reduce wave amplitude from 15mm to less than 1mm within one metre. The radar reflection from such a surface will depend partly on the roughness of the grease ice which is formed but grease ice forming under calm conditions may well satisfy the requirements for coherent reflections.

4.7.3 Surface roughness of newly formed ice

During the BEPERS (Bothnian Experiment in Preparation for ERS-1) campaign, profiles of surface roughness on the mm scale of newly formed ice were collected, using surface based laser profilometry. Table 4.3 shows the the mean roughness parameters given by Johanssen [1988].

The degree to which snow on the surface of the ice observed by Johanssen would affect the backscatter observed by a radar altimeter is unclear. Even adding the roughness of the ice and snow would still lead to a surface with coherent backscatter over the distance of observation (~1 metre). If longer wavelength roughness (which was not measured) was negligible, the resulting backscatter from such ice could be as large as 25-35dB, assuming a Fresnel reflection coefficient of -11dB (as compared with -2dB used to generate the values shown in figure 4.8a).

\[
\sigma_m (\text{mm}) \quad l (\text{mm})
\]

<table>
<thead>
<tr>
<th></th>
<th>(\sigma_m) (mm)</th>
<th>(l) (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Test area ice</td>
<td>2.0±.15</td>
<td>14.0±2.6</td>
</tr>
<tr>
<td>snow</td>
<td>1.6±.84</td>
<td>87.7±28.8</td>
</tr>
<tr>
<td>Intensive area</td>
<td>ice</td>
<td>2.7±.64</td>
</tr>
<tr>
<td>snow</td>
<td>3.9±2.0</td>
<td>85.1±33.9</td>
</tr>
</tbody>
</table>

Table 4.3 Mean surface roughness parameter determined during BEPERS-88 [Johanssen, 1988].

Looking at figures 4.8b and 4.8c it is also evident that the radar returns from such ice would show significant fall off within the range window. In chapter 7 we will show that smooth ice in
the Kara sea, which may be similar in nature to that observed by Johanssen, results in AGC values of between 20-30dB over the smoother areas.

4.7.4 Surface roughness of ice floes

With the exception of the work by Johanssen, most observations of sea ice roughness are carried out using helicopter-borne laser profilometry and ground survey. These techniques cannot provide measurements to the millimeter accuracy needed for frequencies of 13GHz.

Profiles of ice floe roughness over summer Arctic ice, collected during the Lance cruise, are shown in appendix A (Table A.1). The minimum RMS surface slope observed over a smooth first year floe was 0.041. With reference to figures 4.6 and 4.7 this would imply a normal incidence backscattering co-efficient of less than 30 dB, assuming a maximum $|R(0)|^2$ of -4dB, and would not result in a significant fall off in return power within the altimeter range window. It should be noted that small scale slopes are not resolved by such ground survey measurements. The surface slope derived must therefore be considered as a minimum and the backscattering co-efficient a maximum.

4.7.5 Surface roughness of meltponds

In section 4.4.3 we described a model for backscatter resulting from a collection of flat areas on the surface developed by Brown [1982]. He suggests that smooth facets on the surface of ice floes may be responsible for the high levels of backscatter observed in altimeter data over sea ice. As stated before he does not present observations of surface roughness of sea ice which support this hypothesis.

However observations made during the Lance cruise, also described in Appendix A, show that the surface roughness of meltponds were smooth enough to result in coherent returns (Plate A.8). Figure A.4 shows that the size distribution of frozen meltponds observed has a peak at around 1-1.5m. Looking at figure 4.3 we can say that if frozen meltponds represented around 10% of the total area, and if the second moment of radius was around 1m, the peak backscatter observed due to reflections from meltponds would be as large as ~30db (assuming $|R(0)|^2 = -11$dB) with significant fall off in power within the range window.
4.8 Discussion

We have shown that the origin of the peaked returns observed in altimeter data may be the result, either of coherent backscatter from some fraction of the surface, or of quasi-coherent backscatter caused by small surface slopes. Few measurements of the surface roughness of either sea ice or ocean to the mm scales required for input to such models exist. General conclusions can however be drawn. Firstly, measurements which do exist of mature (i.e. greater than a few days old) first-year and multi-year ice floes would seem to indicate that these are unlikely sources for the high power, peaked return waveforms that are typically observed over sea ice. Secondly observations of meltponds and new ice and theoretical models for the ocean surface roughness in leads and open water within the pack, suggest that quasi-coherent or coherent backscatter is certainly possible from such surfaces. These conclusions are supported by comparisons between altimeter and AVHRR data presented in chapter 7.

4.9 Conclusions

We can summarise the results from this chapter as follows:

- The backscattered power from a surface depends on its di-electric properties and on surface roughness characteristics.

- Changing di-electric properties can account for a variation of about 10dB in the backscatter from sea ice. Since observations of backscatter in altimeter data show variations over more than 50dB we must conclude that changes in surface roughness is the primary cause of backscatter variation.

- The backscatter from a perfectly flat surface will result in an effective backscatter around +50dB, some $10^4$ times greater than typically observed over the ocean.

- Over some areas the radar altimeter return may be dominated by reflection from discreet coherent areas on the surface.

- If large flat areas exist on the surface, near field effects may cause problems when comparing airborne and satellite observations.
• Small perturbation theory shows that coherent backscatter become negligible for a surface roughness greater than 6mm.

• Physical optics models show that high levels of backscatter can still exist for surfaces with a large surface roughness if surface slopes are small.

• Two scale models combine the predictions of small perturbation theory and physical optics models. They show that changes in small scale roughness may act to modify the level of backscatter observed independent of any other changes in the nature of the sea ice cover.

• Observations of normal incidence backscatter from sea ice show much lower levels than are typically observed in altimeter data.

• Ice floes damp out water waves, particularly at higher frequencies, but wind can regenerate small scale wave within the ice pack depending on the prevailing wind speed and fetch available.

• Comparing observations of surface roughness in sea ice areas with the predictions of backscattering theories shows that smooth water areas, meltponds and areas of new ice are the most likely sources for the high power peaked returns observed over sea ice, whilst returns from older ice floes is more likely to result in diffuse, low power returns.
CHAPTER 5 DATA PROCESSING

5.0 Introduction

In this chapter we describe the formats for the Seasat and Geosat data presented in later chapters. We then go on to describe the methods of data correction and reduction employed for altimeter data over sea ice.

5.1 Altimeter data formats

5.1.1 Seasat data formats

Telemetry directly received from the radar altimeter on board the Seasat altimeter is recorded in the Sensor Data Record (SDR). Data is recorded at the rate of ten frames per second, each frame being the sum of two 50 pulse waveform averages and associated tracking parameters. A summary of the data appearing in the Seasat SDR is as follows:

(i) Sixty contiguous samples of return power, three further samples around the tracking point and an attitude gate.
(ii) Output of height and gain control loops: Range, SWH, AGC, range rate
(iii) Time/Location information: Time tag, Spacecraft height, location
(iv) Spacecraft health parameters

Once recorded the SDR is processed to produce the Geophysical Data Record (GDR), details of processing are given in Hancock et al. [1980]. Each frame in the GDR is produced from ten frames in the Seasat SDR. A summary of parameters given in the Seasat altimeter GDR is as follows:

(i) One second averages of altimeter derived parameters: Elevation (above the reference ellipsoid), SWH, Backscatter coefficient, Wind speed
(ii) Quality information: One second standard deviation of elevation, SWH and backscatter
(iii) Quality flags: Flags indicating poor data quality and spacecraft health
(iv) Time/Location information: Time tag, Spacecraft height, location
(v) Corrections: Tropospheric and Ionospheric range corrections
(vi) Ancillary information: Geoid height, Tides

The GDR data is intended for use by researchers interested in applications discussed in section 1.5. As discussed in chapter 3, over sea ice and other non-ocean surfaces instrumental problems frequently cause large errors in the values of elevation, SWH and backscattering co-efficient. To reduce these errors requires reprocessing of waveform data from the SDR. The techniques employed to achieve this are the subject of this chapter.

5.1.2 Geosat data formats

Although the altimeter instruments flown on board the Seasat and Geosat satellite are essentially identical the formats in which the data is distributed differ significantly. The main difference is that the waveform data is split from the main Sensor Data Record (SDR) and is distributed as a Waveform Data Record (WDR). Another important difference for users of the data is that location information appears only in the GDR. The Geosat SDR contains data at the rate of one frame per second in contrast to the 10 Hz rate for Seasat SDR data.

Data which appears in the Geosat SDR is as follows:

(i) Output of height and gain control loops: Range (10 samples), SWH, AGC (10 samples), range rate, e.t.c.
(ii) Voltage proportional to attitude parameter (VATT)
(iii) Spacecraft health parameters

Waveform data from the geosat altimeter is also recorded but placed in a separate data set known as the Waveform Data Record (WDR). This contains waveforms samples, similar to those contained within the Seasat SDR, which may be associated with Geosat SDR data by use of a frame counter.

The SDR is processed in a similar manner to Seasat to produce the Geosat GDR. Parameters contained within the geosat GDR are as follows:
(i) Altimeter derived parameters: Initial height, individual height offsets (10 samples), AGC (10 samples), SWH, Backscatter coefficient, Wind speed

(ii) Quality information: Standard deviation of elevation, SWH and Backscatter coefficient.

(iii) Quality flags: Flags indicating poor data quality and spacecraft health

(iv) Time/Location information: Time tag, Spacecraft height, location

(v) Corrections: Tropospheric and Ionospheric range corrections

(vi) Ancillary information: Geoid height, Tides

Data in the Geosat GDR is further split into an 'oceans' and a 'land/ice' GDRs. Details of the criteria by which individual data points are placed in one or other are given in Cheney, et al. [1987].

5.2 Processing of altimeter data over sea ice

The parameterisation of radar altimeter return waveforms employed by Drinkwater (section 2.6.5.5) and Ulander (2.6.5.6) provide some improvement on single ice index parameters employed by other researchers. The waveform parameters chosen for analysis here are similar in some respects and have been chosen specifically to allow discrimination between different scattering mechanisms. In the analysis presented in this work the retrieval of high precision range measurements is an additional goal of algorithms employed to process the data. This aim in particular requires detailed consideration of the instrumental problems discussed in chapter 3 and implementation of quality control where telemetry summing prevents retrieval of accurate range estimates.

The reprocessing of altimeter data over sea ice has two main aims:

(i) To correct for instrumental errors associated with operation over non-ocean surfaces.

(ii) To extract parameters which may be useful over non-ocean surfaces.

In addition some effort was made during analysis to reduce data volume to allow processing and averaging of relatively large data sets at one time.
To achieve these aims a processing scheme was devised to produce a summary file from the Seasat SDR data to be used in later analysis.

The processing steps employed are as follows:

(i) Return waveform quality control
(ii) Waveform sampler correction
(iii) Waveform retracking
(iv) Correction of antenna pattern
(v) Conversion to $\sigma^0$ values
(vi) Waveform parameterisation

Once these processing steps had been carried out a further data set consisting of sea ice summary files was produced.

5.3 Waveform quality control

As outlined in chapter 3 various instrumental effects can cause the waveform data to be corrupted in such a way that retrieval of useful parameters is severely compromised. As part of the processing a quality flag was produced and included in the summary files which indicated whether one or more of the instrumental effects was present. Coding of the quality flag was achieved as by setting it to an initial value of zero and then adding powers of 2 according to which tests the telemetry frame failed. Coding was as follows:

(i) Telemetry missing add 1
(ii) Loss of lock add 2
(iii) Waveform spiking add 4
(iv) Double peaked add 8
(v) Height glitch add 16
(vi) Snagged waveform add 32

During data analysis the quality flag could simply be decoded according to what criteria were required for a data point to be included.
The first two criteria are tested by looking at status words given included in the SDR. Waveform spiking is an instrumental effect of unknown origin which leads to one or more samplers (normally one) having anomalously high count values. Double peaking normally arises from the erroneous addition of two peaked waveform profiles resulting from tracker oscillation. Further details of the algorithms employed for the detection of 'waveform spiking' and 'double peaking' are given in Rapley et al. [1987].

![Figure 5.1](image)

**Figure 5.1** Altimeter data over sea ice showing tracker oscillation and the effect of the 'height glitch' phenomenon. During the rapid increase in elevation, as telemetred, the retracked elevation shows a large error due to problems of telemetry summing. By applying thresholds to the height error, which is also supplied in the telemetry, data which is severely effected by the 'height glitch' problem can be eliminated.

In section 3.4.4 we briefly discussed the 'height glitch' phenomenon which occurs when two altimeter frames are summed during a rapid excursion of the range window. Figure 5.1 shows raw and retracked elevation values over sea ice in the Antarctic. The phenomenon of tracker oscillation is clearly seen in the raw elevation values with a rapid (positive) departure from the true elevation value followed by a gradual fall. The retracked elevation profiles closely follow the raw values during the rise to a maximum error resulting in the 'height glitch' phenomenon. This phenomenon results from the manner in which the telemetry is summed Rapley et al. [1987]. Although the problem of 'height glitches' cannot be completely overcome the effect can be reduced by considering the height error which is used as an input to the height tracking
loop (section 3.3.3). At times of maximum height error the tracking loop range will be driven at the greatest rate resulting in the maximum error in the retracked elevation. Looking at figure 5.1 there is a clear correlation between peaks in the height error value and maximum departures in the retracked elevation value. To flag such occurrences a threshold test on the value of the height error is used. Optimum values for the thresholds employed depend on the return waveform shapes and ideally should be determined by extensive modelling. As a first order correction the thresholds on the height error value, outside which the height error flag was set, were chosen to be -0.5 m (minimum) and +2.0 m (maximum).

Tracker snagging is discussed in Rapley et al. [1983] and results from the altimeter passing over regions of rapidly varying backscatter. Detection of this phenomenon is also problematic. Again first order solution is employed. Typically the leading edges of snagged waveforms consists of a preliminary low power return from the surface directly below nadir and a peak some distance into the waveform depending on the off-nadir range of a bright surface. The empirically derived method of detection is to determine whether the waveform profile is monotonically increasing before the peak sampler is reached. This approach is generally valid for 'peaked waveforms' but cannot be applied to diffuse 'ocean' like waveforms where fading noise may result in several single sample minima before the peak is reached. To overcome this 'ocean' like waveforms, as determined by the pulse peakiness test [Laxon and Rapley, 1987], are excluded from this test.

It should be said that the tests described above are intended primarily to flag data points where the retrieval of precise range measurements is not possible. Although instrument saturation (section 3.4.1) causes some problems when waveforms become highly peaked, it was decided that filtering such data from analysis presented later might lead to the exclusion of potentially interesting data. For this reason no flag is set to indicate instrument saturation although such a check should be employed where the retrieval of precise (within a few dB) backscatter estimates is important.

5.4 Waveform sampler gain

To obtain an approximate correction for sampler gain, a waveform average over a section of ocean data, where the variation in pointing angle and significant wave height was minimal, was obtained. The resulting mean Seasat waveform is shown in fig 5.2 along with a model
waveform generated using Browns model given in section 3.6.1 for an equivalent SWH and off-pointing angle.

![Waveform graph](image)

Figure 5.2 Seasat waveform sampler gains. The two profiles shown are the mean observed waveform for a section of data over the open ocean and the theoretical waveform for an equivalent off-pointing angle and SWH derived from equation 3.12.

By taking the ratio of the two waveform profiles an approximate gain correction value for each sampler was determined. These gain values were then applied to all Seasat waveform data before parameterisation was carried out. The correction of Geosat waveforms was carried out using sampler gains provided in the header information of the WDR tapes.

5.5 Waveform retracking

The operation of the height loop on board the Seasat altimeter generally caused large errors to appear in the telemetred range value when operating over non-ocean surfaces. A significant reduction of this error can be made using a method known as waveform retracking. This method involves identifying the point on the return waveform that corresponds to the mean surface level. The offset between this point and the centre of the range window, to which the predicted range measurement is made, can be used to calculate a correction to the range measurement. The approximate location of the first return is also required to allow consistent waveform parameterisation in terms of off-nadir angle.

Several different methods of retracking have been employed by researchers in the past. Waveforms over land ice were retracked using a least squares model fit [Martin et al., 1983]. Zwally et al. [1987b] also used the retracking technique devised by Martin, but
used operator editing to cope with poor fits to more peaked return waveforms. Laxon and Rapley [1986] used a least squares fitting technique to model templates of sea ice returns to retrack data over sea ice with some success. The model used by Martin is however not ideal and more recent retracking of land ice returns has been carried out using simple thresholding techniques [Partington, 1988].

![Figure 5.3: Threshold retracking of altimeter waveforms.](image)

The point on the return waveform corresponding to the mean surface level is identified to be at a point \( x \) bins into the range window. The range offset \( \Delta h \) is simply calculated by multiplying the bin offset by a scaling factor thus:

\[
\Delta h = \frac{x}{2c} (x-30.5) = 0.47(x-30.5) \text{ metres}
\]

5.1

Rather than use a simple threshold on \( x \) to identify the value of \( x \), a linear interpolation is used between the two bins which lie either side of the 50% peak power point.

5.6 Retrieval of backscattering co-efficient from Seasat data

We start by substituting the parameters for the Seasat altimeter into the link equation for the case of a pulse limited altimeter. Values used are \( G = 40.8 \text{dB}, \ h_0=8.10^5 \text{m}, \ \lambda=2.2 \text{cm} \ c=3.10^8 \text{ms}^{-1}, \ \tau = 3.125.10^{-9} \text{s}. \) A value of \( L_{\text{atm}} = -1 \text{dB} \) is assumed although this may be one or two dB more depending on atmospheric conditions. Substituting in equation 4.5 we obtain:

---

128
From the system parameters for the Seasat altimeter the value for $P_r/P_t$ for any sampler, $i$, can be obtained from:

$$P_r = \text{AGC} - 198.33 + 10\log_{10}(P(i)) \text{ dB}$$

therefore $\sigma^0$ for any waveform sampler is:

$$\sigma^0 = \text{AGC} - 40.43 + 10\log_{10}P_i \text{ dB} + 10\log_{10}\left[\frac{h^3}{h_0^3}\right]$$

The extraction of backscatter levels for Geosat waveforms is similar, but with a different constant value substituted for the value of 40.43 used in equation 5.4.

### 5.7 Sea Ice waveform parameterisation

Ideally the reduction of waveform data should be carried out by fitting some sort of parameter model using a least squares fit. This method is used by Martin et al. [1983] who developed a model for altimeter returns from continental ice sheets. However, no single model has yet been demonstrated as being suitable for altimeter return waveforms over sea ice. As an alternative parameters derived from the altimeter waveform profiles should be chosen to retain information about key characteristics of the waveform shape.

The selection of parameters has been a somewhat iterative process. The main aims of the parameterisation are to distinguish, if possible, between the different types of return that occur and identify those which resemble pulses predicted by backscattering theories discussed in chapter 4. The waveform parameters are grouped as follows:
(i) Leading edge parameters - giving information concerning the surface roughness.
(ii) Peak backscatter value - giving information on the di-electric properties and
normal incidence radar reflection.
(iii) Trailing edge parameters - giving information on the backscattering theory most
applicable to the data and quantitative information on the surface roughness
characteristics.

5.7.1 Leading edge parameters

For a diffuse return the leading edge yields information about the meso-scale surface
roughness. In particular the differential of a return from a diffuse surface, after correction for
antenna and waveform sampler gain, is the probability distribution of point scatters on the
surface. Over the ocean this distribution is normally very close to a Gaussian and therefore
requires only one parameter the standard deviation $\sigma_h$ ($\sigma_h = 4xSWH$).

Over sea ice two physical characteristics are likely to cause the return pulse shape to deviate
from a Gaussian distribution. Penetration will cause the distribution to be skewed below the
median [Ridely et al., 1987] whilst the appearance of ridges will cause the distribution to
be skewed above the median.

![figure](image.png)

**Figure 5.4** Calculation of leading edge parameters

Three points on the leading edge of the return waveform are defined where the pulse profile
reaches some fraction $T$ of the peak value. The position of each point is found from linear
interpolation using:

$$x = (i - 1) + \frac{T - P_{i-1}}{P_i + P_{i-1}}$$

5.5
where the \( i \)th bin is the first sample on the leading edge to exceed \( T \). Threshold chosen during the analysis of the Seasat data were:

\[
\begin{align*}
T_{\text{LOW}} &= 0.25 \times \text{PEAK} \\
T_{\text{MED}} &= 0.5 \times \text{PEAK} \\
T_{\text{HIGH}} &= 0.75 \times \text{PEAK}
\end{align*}
\]

where \( \text{PEAK} \) is the maximum count value in the waveform. Two leading edge parameters are calculated from the three retrack points.

The \( \text{LEWID} \) parameter provides a measure of the leading edge width and is calculated using:

\[
\text{LEWID} = \frac{(x_{\text{HIGH}} - x_{\text{LOW}}) \times ct}{2}
\]

The second parameter \( \text{LEDIF} \) provides a measure of the skewness of the leading edge and is calculated using:

\[
\text{LEDIF} = \frac{(x_{\text{HIGH}} + x_{\text{LOW}})}{2} - x_{\text{MED}}
\]

**Figure 5.5** Leading edge parameter calculation for different surface height probability density functions.

Figure 5.5 shows how the values of \( x_{\text{LOW}}, x_{\text{MED}} \) and \( x_{\text{HIGH}} \) are affected by a distortion of the height distribution of point scatterers. We can distinguish between no distortion or distortion caused by penetration or ridges. Roughly speaking for each class we have:
5.7.2 Peak backscatter value

The peak backscatter value, SIGPK, under conditions of homogeneous ice cover, represents a measure of the backscattering coefficient at normal incidence. This measure can be directly related to the predictions made for the backscatter theories discussed in chapter 4. Its value is computed using equation 5.4 with $P_i$ equal to the peak count value in the return waveform.

5.7.3 Trailing edge parameters

The main change that occurs in the trailing edge of the return when the altimeter observes sea ice is to waveforms where a rapid fall off in return power occurs away from the nadir point. The first measure adopted is therefore a measure of the magnitude of this fall off by computing the power in two gates, one at the nadir point and one near the end of the range window. The difference between the power in these two gates is called SIGTD. A second measure is also adopted to measure the deviation of the trailing edge of a return waveform from an exponential fall off. This will allow detection of the point at which the return waveform starts to deviate from the Quasi-specular stationary phase approximation (Gaussian slope distribution) which is the theory normally employed over the ocean (See chapter 4). Processing therefore starts by defining a leading edge bin, LEB, which lies at or just after the peak return in the waveform. Three 'broad sea ice gates are then calculated as follows:

$$\text{SIGT0} = \log_{10} \left[ \sum_{i=\text{LEB}}^{\text{LEB+7}} P_i \right] + K$$

$$\text{SIGT1} = \log_{10} \left[ \sum_{i=\text{LEB+8}}^{\text{LEB+15}} P_i \right] + K$$

$$\text{SIGT2} = \log_{10} \left[ \sum_{i=\text{LEB+16}}^{\text{LEB+23}} P_i \right] + K$$
Where $K$ is an overall gain value obtained from equation 5.4:

$$K = AGC - 40.43 + 10 \log_{10} \left[ \frac{h^3}{h_0^3} \right]$$  

These three gate values are stored in the sea ice summary file.

**Figure 5.6** Three sea ice gates used to characterise the trailing edge of the return waveform.

Two further parameters, $SIGTD$ and $SIGTR$, are calculated from the three broad gates using:

$$SIGTD = SIGT0 - SIGT2$$  

$$SIGTR = \frac{SIGT0 + SIGT2}{2} - SIGT1$$

These secondary parameters are widely used in the analysis presented in later chapters. The choice of parameters is also intended to be suitable for processing data over sea ice from the radar altimeter to be flown on board ERS-1. The different operational modes of ERS-1 mean that range bins may be either ~3ns ('oceans mode') or ~12ns ('ice mode') wide. When the ERS-1 altimeter is in 'ice mode' the average of 2 instead of 8 samples will be used to compute the broad sea ice gates allowing mode switches to be transparent in the sea ice products which have been specified.
Two types of data format are available from the Seasat and Geosat altimeter instruments. Sensor Data Records (SDR's) contain on return waveform profiles and tracking parameters at 10Hz along with details of instrument operating modes. Geophysical Data Records (GDR's) contain geophysical parameters at 1Hz and include ancillary information such as atmospheric corrections, geoid and tidal values. Analysis of the SDR's is necessary to overcome instrumental problems which occur over sea ice.

Quality control is implemented to eliminate data where retrieval of high precision elevation data is not possible.

A novel approach to parameterisation is used based on predictions of waveform shapes from backscattering theories and knowledge of the likely surface roughness characteristics in sea ice covered ocean areas.
6.0 Introduction

In this chapter we present analysis of data over sea ice from the Seasat altimeter. We begin by demonstrating that the parameters described in chapter 5 delineate distinct zones within the sea ice cover off East Greenland. A study of the temporal variation in altimeter derived backscatter is carried out using data from the 3 day, near repeat cycle. In the final section we present an analysis of Seasat altimeter data in the Northern and Southern hemispheres for a complete 17 day repeat period. The data set used forms part of the Polar Reference Data Set (PRDS) extracted by the Algorithm Development Facility (ADF) based at the Royal Aerospace Establishment (RAE) in Farnborough.

6.1 The Seasat mission

The Seasat satellite was launched on the 27th June 1978. After an initial commissioning period, the altimeter was placed in a 17 day repeat orbit. Around the 14th of September 1978 the orbit was changed so that the satellite performed a repeat cycle every 3 days. The satellite continued in this orbit until it finally failed on the 10th of October 1978.

The timing of the Seasat mission corresponds to mid-summer and early winter in the Arctic and to mid-winter and early spring in the Antarctic. Differences in the altimeter response over the Arctic and Antarctic regions may, therefore, be due to difference both in the nature of Arctic and Antarctic sea ice cover and in seasonal changes.

6.2 Altimeter parameters over the East Greenland Sea

To begin our analysis of satellite radar altimeter data we consider the behaviour of the waveform parameters outlined in chapter 5 for a segment of data from East Greenland. This data is selected because the altimeter parameters show excellent delineation of apparently different zones of backscattering characteristics.
Figure 6.1 Waveform parameters for a transect off the East Greenland coast. The waveform parameters (SIGPK, SIGTD and SIGTR) are observed to clearly delineate four different zones from the ice boundary inwards. Excursions in the value of LEWID are observed at the first two boundaries due to tracker 'snagging'. Each block corresponds to approximately 15km of ground track.
Figure 6.1 shows linear time line plots of four waveform parameters; SIGPK, SIGTD, SIGTR and LEWID. Each horizontal tick mark corresponds to five blocks or approximately 75km or 11 seconds of altimeter track. Open ocean values of SIGPK and LEWID are approximately 15dB and 1m respectively. This indicates that conditions were extremely calm with a low wind speed. Several distinct transitions occur as the altimeter passes from open water to the area of sea ice cover indicated by rapid increases in one or more of the three waveform parameters, SIGPK, SIGTD and SIGTR.

Zone 1 extends over some 35 km with a rapid rise in the value of SIGPK followed by a more gradual rise. Little change occurs in the values of SIGTD and SIGTR indicating the reflections from the surface are still essentially isotropic in the narrow range of sampling angles afforded by the altimeter. Referring to the discussion chapter 4, the initial rise in SIGPK may be caused by damping of wind waves, whilst the more gradual rise which follows may be caused by damping of longer wavelength swell.

Zone 2 begins with a rapid rise in SIGTD, by about 12 dB, and a slight increase in the value of SIGTR. to around 1-2 dB, indicating a transition from ocean like to very peaked waveforms. Beyond this point, a fairly constant value in SIGPK, of around 45-47 dB, is observed. This may result from an area of fairly constant ice cover where regeneration of wind waves between floes is maintained at a constant level, preventing a transition to purely specular returns.

Waveforms in this section closely resemble the predictions of the specular point theory stationary phase approximation with a Gaussian distribution of surface heights (see section 4.5.3). A least squares fit to the logarithm of return waveform profiles in this area was used to determine their 3dB width. Using equations 4.28 and 3.5 the effective reflection coefficient, the product between the Fresnel reflection coefficient and the fraction of the surface, F, responsible for quasi-specular returns, was calculated using:

$$|F \cdot R(0)|^2 = s^2 \sigma^0(0^\circ) = \frac{W_{3B}}{666200} \cdot SIGPK$$

Figure 6.2 shows a plot of $F \cdot R(0)^2$ for a sequence of pulses around block 1210.
The values of $|\mathbf{R}(0)|^2$ obtained lie between 0.15 (-8.3dB) and 0.35 (-4.5dB), significantly higher than the theoretical maximum for a 100% ice covered surface. We can therefore say, with some certainty, that water between ice floes is responsible for the peaked, high power returns that are observed. It may appear, at first, that the factor $F$ gives us a measure of the percentage of open water and hence of ice concentration. However the difference between the observed backscatter and that which would result from 100% water may equally well be explained by superimposed small scale roughness which would attenuate the peak backscatter as predicted by the two scale model (section 4.5.3).

The start of zone 3 is delineated by a marked increase in the value of SIGTR suggesting a change in the scattering mechanism at this point. This may represent a transition to a surface where the specular component becomes important as the surface roughness falls below 6mm.

The last section, zone 4, shows a lower and more variable SIGPK value indicating a less homogeneous ice cover on scales of the altimeter footprint. This may be indicative of an ice regime with a population of large floes, interspersed by smaller floes and areas of open water. Two strong dips are observed in the SIGPK profile, accompanied by a drop in SIGTD values which indicate diffuse returns. Figure 6.3 shows a Landsat RBV image obtained 9 hours after Seasat passed over the same area. The image covers an area of 99x99 km corresponding to a scale of approximately 2mm to 1km.
Figure 6.3 Final section of the Seasat track over the East Greenland Sea shown in figure 6.1. The altimeter pass took place at approximately 0400 and the image was acquired at approximately 1300 GMT on the 22nd of August 1978. The image shows the general ice regime to be of low concentration but with numerous vast floes present close to the altimeter track.
Errors in the location of the altimeter track, and by movement which has occurred since the time of the altimeter pass and image acquisition make direct comparison difficult. However, the overall regime is seen to be populated by vast ice floes of 10-20 km across with smaller floes and open water in between. Such vast floes are primary candidates for the low power ocean like returns, indicated by the two dips in the SIGPK profile and corresponding fall in SIGTD values.

This analysis demonstrates that distinct zones appear in profiles of altimeter parameters over sea ice. The spatial scale of zones observed, around 30km in length, is comparable to the scale zones described in chapter 2 (section 2.2.2), although the sea ice regime on this coast is not typical due to the influence of the East Greenland current. Correlation of zones observed in the altimeter data and those which can be related to sea ice properties requires comparison with ground observations or data from other sensors.

6.3 Temporal variations in altimeter data over sea ice

Having looked at the spatial variation in the altimeter signal over a single track we now examine temporal variations in the altimeter signal over sea ice. Data from the three day repeat cycle is presented for a single track over the Southern ocean shown in figure 6.4. Temporal variations may be caused both by changing sea ice cover or by differing wind or wave conditions.

![Figure 6.4](image)

*Figure 6.4* Location of the ground track used in the three day repeat cycle study.
Figure 6.5 shows linear plots of the peak backscatter value (SIGPK) at intervals of 3 days for the period 18th-27th of September 1978. The inward track shows a well defined edge followed by a steady increase in backscatter from ~20 to ~40 dB over some 300 km. The shape of the backscatter profile is similar for different passes but there is considerable movement in the position of the initial rise. After the initial rise there is a sharp drop where the backscatter falls to around 15 dB. The level of backscatter then shows some recovery although now there are significant differences in the profiles for the separate tracks. The outward pass start with a section of low backscatter, around 0dB, close to the coast extending over some 200 km. This is followed by an area of higher backscatter between ~20 to ~40 dB over some 900 km. In contrast to the inward track, backscatter profiles in this area show large differences between different passes. Whether the observed changes are caused by changing ice or weather conditions is not clear but the low backscatter returns at the start of the outward pass deserve closer examination.

Figure 6.6 shows the waveform sequences for the four different passes close to the edge of the low backscatter area some 100 km from the coast. Waveforms with a peak backscatter greater than 7dB have been normalised to the same maximum value and are indicated by shading of the peaks. Waveforms for the 18th and 21st of September appear to be diffuse returns with a number of superimposed peaks. The low level of peak backscatter indicates that flat areas, making up a very small percentage of the total surface area, are probably responsible. Waveforms for the 24th and 27th of September are of lower power show fewer peaks. Bright features do however remain at the beginning and end of the sequence. Over these regions the waveform shapes show remarkable similarity between passes.

These observations show that altimeter data over sea ice exhibits remarkable temporal stability in some areas whilst showing large variations in others. The diffuse returns observed at the beginning of the outward pass show a level of backscatter much lower than is normally observed over the ocean. As with the East Greenland data it seems most likely that such returns originate from an area of 100% ice cover. One possible explanation of the changes observed is that we are observing an area where the ice floes are converging, closing up small leads which are causing the more peaked returns observed in the 18/9 and 21/9 passes. This is not unreasonable since the observations are close to the coast and may lie in the shear zone described in chapter 2.
Figure 6.5 Profiles of peak backscatter (SIGPK) for four repeat cycles over the altimeter track shown in figure 6.4. Tick marks correspond to 660km of ground track. Backscatter profiles on the inward pass show a consistent rise in backscatter at increasing distance from the ice edge. Backscatter profiles on the outward pass are more variable over the bulk of the sea ice but show consistent, low backscatter values near the coast.
Figure 6.6 Waveform profiles for four different repeat cycles near the coast on the outward passage of the 3 day repeat track.
6.4 Distribution of waveform parameters values from the Polar Reference Data Set

In this section we examine the spatial variations in altimeter waveform parameters observed over a seventeen day period. Before plotting data has been averaged into geographical bins each covering 0.4° of latitude and 2° of longitude to aid clarity.

6.4.1 Distribution of altimeter parameters in the Antarctic

Average values of waveform parameters for the southern hemisphere are shown in figure 6.7. The scale at the bottom of the image shows that lowest values are plotted in mauve whilst the highest are plotted in red and white.

On the periphery of the SIGPK plot values are first seen to increase from the open ocean value to a slightly higher value, indicated by a transition to light blue, for one or two bins (corresponding to 40-80km). This transition seems to come before a significant change in the pulse shape occurs, since SIGTD values increase at some distance in from the initial change in SIGPK. This zone would seem to represent an initial damping of ocean wave roughness by ice floes or by new ice forming, as observed previously in section 6.2, resulting in diffuse (or 'ocean like') returns with higher backscatter levels than seen over the ocean. This zone corresponds with passive microwave measurements of ice concentrations of between 15 and 50%. Within this the higher backscatter values are observed, in a zone with a width by between 100 and 200 km around the continent, which, referring to figure 2.5b, corresponds to PM ice concentrations near the 50% ice boundary. Further into the ice pack, lower values of backscatter are observed, corresponding with ice concentrations above approximately 80%.

At this point we might postulate that, near the ice boundary, regeneration of wind waves, owing to the larger inter-floe distance, and swell penetration prevents quasi-specular reflection. Far into the ice pack, although smooth water surfaces may exist, their fractional coverage will be smaller. The zone of highest backscatter, which lies between these two, lower backscatter zones, represents a balance between these two effects.
Figure 6.7a Antarctic SIGPK values. Highest values of peak backscatter are observed just within the ice edge. This probably represents the point at which the maximum coverage of open water exists whilst the inter-floe distance remains small enough to prevent regeneration of wind waves.

Figure 6.7b Antarctic SIGTD values. Highly peaked waveforms tend to correspond to the highest SIGPK values observed in figure 6.7a. Inside this zone area peaked waveforms occur less frequently, possibly due to the lower probability of open water areas being present.

Figure 6.7c Antarctic SIGTR values. The values of SIGTR observed over the southern ocean suggest that few coherent returns exist and that peaked waveforms are mainly the result of quasi-coherent backscatter.

Figure 6.7d Antarctic LEWID values. The LEWID parameters provides an estimate of the metre scale surface roughness. Values over the open ocean are highest due to swell storms. Within some sea ice areas values greater than 2 metres are observed but these may be caused in part by tracker snagging. A large area of high LEWID values is also observed off the Oates coast in a region where rough multi-year ice is expected.

Figure 6.7 Averaged waveform parameter plots for the Southern hemisphere
It is also worth noting that the lowest levels of backscatter, within the pack, are observed in areas where sea ice cover exists throughout the year. A good example of this is the area off the Oates coast (150-160°E, 68°S) which shows a large region of low backscatter.

The distribution of SIGTD values, an indicator of how peaked return waveforms are, shows different patterns to those seen in the SIGPK values. High values are seen up to the coast in most regions, except the Weddell and Belinghausen seas, where large regions of multi-year ice may be expected. Again the area off the Oates coast is distinguished as an area of diffuse returns. North of Queen Maud land (20°W-20°W, 70°S), an area of very peaked returns is observed. This corresponds with an region where significant shoreline leads were observed during a cruise by the Polarstern in 1986 Eicken and Lange [1989].

Values of SIGTR show a maximum, around the inner boundary of the high SIGPK values around the periphery, falling to a lower value further into the pack. Average values, at around 1-2dB, are however somewhat lower than observed in the zones described in section 6.2, and may not necessarily indicate a change in backscatter mechanism.

Variations in the leading edge width parameter, LEWID, may be caused, either by significant surface roughness, in the case of well behaved diffuse returns, or by areas of large (range Window Footprint) scale variations in the backscattering of the sea ice cover. The latter is probably the cause of higher values observed in the Weddell and Belinghausen Seas. Over the ocean, variations reveal the different sea states which existed over the seventeen day period. At the ice boundary, a sharp fall in LEWID, is generally observed, indicating that swell penetration is limited to only a few kilometres. Looking more closely at the Oates coast areas, high values of LEWID are observed over the region of diffuse returns. Results, presented later (section 7.4) suggest that this type of return is only observed over old multi-year ice, and the variations in LEWID may reflect variations in the metre scale roughness.
Figure 6.8a Arctic SIGPK values. Open ocean values are higher than is observed in the southern ocean due to the calmer conditions. Values over sea ice are also significantly higher than those observed over Southern ocean sea ice.

Figure 6.8b Arctic SIGTD values. Values of SIGTD show that a higher percentage of highly peaked waveforms are observed in the Arctic as compared with the Antarctic. It is also noticeable that the highest values (indicated by red) are observed some way into the ice pack.

Figure 6.8c Arctic SIGTR values. High values in the SIGTR parameter show that a higher percentage of coherent returns are observed in the Arctic as compared with the Antarctic.

Figure 6.8 Averaged waveform parameter plots for the Northern hemisphere.
6.4.2 Distribution of altimeter parameters in the Arctic

Figure 6.8 shows the distribution of altimeter parameters in the Arctic. The area of sea ice lying within the coverage of Seasat is much less than is seen for the Antarctic. Delineation of zones mapped by altimeter parameters is therefore much less obvious. The main areas where sea ice cover is observed are the East Greenland coast, the Canadian Archipelago, the Chukchi Sea and east of the Kara peninsular. Over the open ocean, peak backscatter values are somewhat higher than in the Antarctic, indicating that prevailing wind speeds are somewhat lower in the Northern hemisphere. Peak backscatter values are significantly higher than those observed in the Antarctic. Lower overall wind speeds would imply less regeneration of wind waves within the ice pack. Also, in contrast to the Antarctic observations, climatological data [Zwally et al., 1987a] show mean air temperatures to be above zero. Values of SIGTD show consistently higher values than are observed for the Antarctic. The highest values are seen to lie some distance away from coasts. Values for the SIGTR parameter are also much greater than those seen in the Antarctic. This may be caused by a higher percentage of returns where a coherent component exists due to reflections from very flat areas of open water or by the presence of meltponds on the ice surface.

6.5 Sea Ice Global backscatter statistics

Figure 6.9 shows histograms of SIGPK derived from the PRDS for the Arctic and Antarctic regions.

Both histograms show a sharp peak between 10 to 15dB due to the inclusion of data over the open ocean. The peak for the Antarctic shows a slightly lower value than for the Arctic probably due to the generally higher sea states present during the southern winter. Some data over land is also be present and may some low backscatter observations.

This result demonstrates that high levels of backscatter exist in the majority of radar altimeter observations of sea ice. The range of backscatter levels is also extremely large, covering at least 50 dB. Two important conclusions stem from this result. Firstly, comparing with the results described in section 4.6, the levels of normal incidence backscatter observed in altimeter data over sea ice are generally much higher than those observed by airborne or ground based
sensors. Secondly the range of variation shows that the variation due to changes in di-electric properties can account for only a small part of the overall variation observed. Changes in surface roughness must therefore be the dominant factor.

![Histogram of SIGPK for Arctic and Antarctic data](image)

**Figure 6.9** Histogram of SIGPK for Arctic and Antarctic data

### 6.6 Conclusions

In this chapter analysis of different Seasat data sets has been presented. The waveform parameterisation shows clear delineation of different zones for a pass in East Greenland and values reflect what might be expected from physical considerations. Data from the three day repeat cycle have proved useful for observation short term temporal variations in the altimeter data. In some areas backscattering coefficients are very similar for consecutive passes while in others large variations are observed. Near the coast the three day repeat data showed a remarkable stability, suggesting that the altimeter is observing areas of quasi-stationary coastal ice or fast ice.

The geographical distributions of waveform parameters shows coherent zones within the Antarctic ice pack. The highest levels of backscatter and most peaked waveforms are observed at intermediate ice concentrations around 50%. Lower levels are observed further into the ice pack where a more compact ice regime is expected. Waveform over the summer
Arctic ice show a much higher level of backscatter in some regions and are also more peaked. This may be due to a greater degree of ponding in the Arctic, as observed during the Lance cruise (Appendix B) leading to coherent returns more frequently.

A histogram of peak backscatter values over a 17 day period show a variation over some 5 orders of magnitude, demonstrating that surface roughness must be the primary cause of variation.
7.0 Introduction

In this chapter we present data from the Geosat altimeter in comparison with data from AVHRR instruments from the NOAA satellite series. The resolution and coverage of AVHRR data is ideal for comparison with radar altimetry. Geosat altimeter data and co-incident imagery are presented for the Kara sea, East Greenland coast and Chukchi sea in the Arctic and for the Weddel sea in the Antarctic. In addition airborne SAR data collected during the Marginal Ice Zone Experiment (MIZEX-87) is compared with altimeter data. Comparisons reveal correlations between altimeter and AVHRR observations of the sea ice boundary, thin ice, leads and fast ice. Vast multi-year floes identified in the imagery correspond to diffuse low power returns in altimeter profiles.

7.1 AVHRR Images used for comparison

AVHRR data presented here was supplied by the Navy Oceans Research and Development Activity (NORDA). Each image is 1024x1024 pixels in size and has been warped to a polar stereographic projection. Data from channel 4 (Infra-red) was available for all images and channel 2 (Visible) data was also used where sufficient daylight permitted its use.

Table 7.1 gives details of the images presented in this chapter for comparison with Geosat altimetry. Centre co-ordinates and scale are given for each image, along with the data and time of acquisition. In some cases WDR (see section 5.12) data was not available, permitting comparison of AGC and VATT values alone. Cloud cover or 'loss of lock' prevented comparison for some images which are not presented.

Note that in all plates presented in this chapter the altimeter moves from right to left with increasing time, whilst the timeline plots of altimeter parameters shown in figures 7.1-7.3 move from left to right with increasing time.
Table 7.1  Regions of comparison for Geosat altimeter data with AVHRR

As mentioned previously (see section 2.6.1) channel 2 data provides the best discrimination between ocean and sea ice during the summer whilst channel 4 data provides the best contrast during the winter period. Most of the images presented were acquired during winter and channel 4 data is therefore presented. The image centre co-ordinates specified in table 7.1 refer to the complete, 1024x1024 image. Images shown in the plates in this chapter are subscenes of 512x512 pixels, except for plates 7.9 and 7.10 which are 256x256 pixels.

For the infra red data pixel values correspond to brightness temperatures $T_\text{b}$ through equation 7.1 [NORDA, personal communication].

$$T_\text{b} = \frac{\text{Pixel value} - 500}{10} \degree \text{C}$$

This relationship is only approximate as no correction for atmospheric effects is made. Images were displayed on a VAX GPX workstation as either sub-sampled (2 image pixels per screen pixel), normal (1 screen pixel per image pixel) or super-sampled (1 image pixel per 2 screen pixels). In all thermal infra-red images (channel 4) images shown in this chapter image intensities have been inverted so that darker areas on the image represent warmer areas on the surface. Contrast stretching has also been performed to allow optimum discrimination between sea ice and ocean surfaces.

<table>
<thead>
<tr>
<th>Region</th>
<th>Image Centre</th>
<th>Date</th>
<th>UT</th>
<th>WDR</th>
</tr>
</thead>
<tbody>
<tr>
<td>South Greenland Sea</td>
<td>67N 30W</td>
<td>10-May-87</td>
<td>15:48</td>
<td>y</td>
</tr>
<tr>
<td>Greenland Sea</td>
<td>74N 15W</td>
<td>18-Mar-87</td>
<td>18:10</td>
<td>y</td>
</tr>
<tr>
<td>&quot;</td>
<td>&quot;</td>
<td>14-Dec-87</td>
<td>18:10</td>
<td>y</td>
</tr>
<tr>
<td>Kara Sea</td>
<td>76N 55E</td>
<td>21-Feb-86</td>
<td>-</td>
<td>n</td>
</tr>
<tr>
<td>Chukchi sea</td>
<td>70N 168W</td>
<td>10-Feb-88</td>
<td>01:32</td>
<td>y</td>
</tr>
<tr>
<td>Chukchi Sea</td>
<td>&quot;</td>
<td>14-Mar-88</td>
<td>16:11</td>
<td>y</td>
</tr>
</tbody>
</table>
7.2 Observations of compact and diffuse ice edges in altimeter and AVHRR data

The accurate determination of the sea ice ocean edge is a major goal of satellite radar altimeter observations. Few direct comparisons have been made of the sea ice edge determined by satellite radar altimetry and by other sensors. In chapter 4 the discussion showed that damping of ocean waves by ice floes was the most likely cause of the sharp increase in return power, at least for areas with less than 100% ice concentration. In such cases the response of the altimeter will not be solely dependent on an increase in ice concentration, but will also depend on the amplitude of the incident swell and the prevailing wind speed.

For a reasonably compact ice edge, such as that studied by Wadhams et al. [1986], a rapid decrease in swell occurs, as the observations presented in section 4.7.1.1 show. If the distance between floes is small, regeneration of wind waves, sufficient to produce rough surface scattering, may not occur depending on the prevailing wind conditions (section 4.7.1.2). Where an ice edge is more diffuse the situation is less clear.

Plate 7.1 shows the Geosat ice index (section 2.4.4.9) compared with an AVHRR channel 2 image off the coast of East Greenland (from Hawkins and Lybanon [1989]). Good agreement between the ice edge observed in the imagery and ice index occurs on all the passes shown. Some structure is observed in the ice index value but it is hard to associate this with changes in the sea ice cover as observed in the imagery.

Plate 7.2 shows the same image, but this time with two waveform parameters, SIGPK and SIGTD (described in chapter 5) overlaid. The SIGTD parameter, an indicator of changing waveform shape, increases from a value of zero over the open ocean to non-zero values close to the ice edge. As expected there is a close correlation between the edge as mapped by the SIGTD parameter and that mapped by the Geosat ice index, which is mainly sensitive to changes in waveform shape. Looking at the SIGPK parameter however the transition from open ocean values occurs some way outside the ice edge, particularly on tracks B and C. This transition appears to occur some way outside the ice edge as determined from the imagery. The cause of this early rise in backscatter is not known, but it may represent the damping of ocean capillary waves by a layer of meltwater (As observed in Plate A.1). Another possibility is
Plate 7.1 Geosat Ice index values over East Greenland [Hawkins and Lybanon, 1989].

Plate 7.2 AVHRR channel 2 image of the Greenland sea 10th May 1987 15:48 SIGPK and SIGTD parameters overlayed.
Plate 7.3  AVHRR channel 4 image of the Greenland sea 18th March 1987 18:10, Geosat ice index overlayed.

Plate 7.4  AVHRR channel 4 image of the Greenland sea 18th March 1987 18:10, SIGPK and SIGTD parameters overlayed.
that if an off-ice wind is blowing, some distance is required before the wind generated waves on the ocean surface reach an equilibrium state.

Plate 7.3 shows the Geosat ice index over a more diffuse ice edge on the 18th of March, 1987. This time the transition over the ice edge visible in the imagery has a very marginal effect on the value of the ice index. Given this plot of ice index values alone, it is probable that the ice edge would be mapped at some distance closer to the coast than its true position. Plate 7.4 shows the values of SIGPK and SIGTD for the same image. Although, as with the ice index, no clear edge is delineated, the value of SIGPK rises to a level well above the open ocean value as it crosses the diffuse ice edge. Less significant changes in the value of SIGTD are observed, although some change from the open ocean values are observed. The more subtle response of the altimeter to the diffuse ice edge most probably results from a longer attenuation distance of swell (see section 4.7.1.1) and a greater opportunity for the regeneration of wind waves in areas of open water due to the greater inter-floe distance as compared with the 10th of May ice edge (see section 4.7.1.2).

To explore this problem further we examine another example of altimeter observations of a diffuse ice edge. Plate 7.5 shows an AVHRR channel 4 image of the ice edge in the East Greenland sea (at a slightly higher latitude than the previous image) for the 14th of December, 1987. The backscattered power as the altimeter traverses from the open ocean into the diffuse ice area is more gradual than was observed previously. Both SIGPK and SIGTD show dips after the initial increase which correspond closely to areas of open water several kilometres wide at some distance into the ice pack. The close correspondence between area of open water, indicated by high brightness temperatures in the AVHRR intensity profile, and dips in the value of SIGPK supports the idea that regeneration of winds occurs in such leads resulting in the less marked altimeter response to diffuse ice boundaries.

The question of the altimeter response to diffuse ice edges is obviously one which must be pursued further to obtain more quantitative estimates of the errors in ice edge determination by satellite radar altimeters. These two examples do, however, show that the compactness of ice boundaries is reflected in the behaviour of the altimeter signals.
Plate 7.5 AVHRR Channel 4 image of the Greenland sea 14th December 1987 18:10, SIGPK and SIGTD parameters overlayed.

Figure 7.1 AVHRR and altimeter line profiles over the Greenland sea 14/12/87. Areas indicated by arrows show regions of open water identified in the AVHRR image, which correlate with dips in the value of SIGPK and SIGTD. This may be caused by regeneration of wind waves in areas of open water several km wide, leading to a return to lower power, 'ocean' like returns.
7.3 Observations of new and compact ice in altimeter and AVHRR data

7.3.1 Kara sea image

Plate 7.6 shows an AVHRR image obtained over the Kara sea on the 21st February 1986. To the east of the Kara peninsula (the land mass observed to run down the left hand side of the image) a large lead has opened between the land and an older area of land ice and subsequently refrozen creating a large area of ice which is perhaps a few days old. The main ice pack shows numerous leads including one large lead between the pack and the fast ice east of the 65°E meridian. The darker appearance of the lead on the Eastern side of the image indicates that brightness temperatures are somewhat higher than for the open area east of the peninsular, suggesting that is is thinner and has formed more recently. Temperatures in this region during mid-winter are likely to be well below freezing and it is reasonable to assume that areas of open water that open up would rapidly freeze over.

Three altimeter tracks are shown on the image. One second averages of the AGC and VATT parameters are plotted vertically from the point of observation. Two primary colours are used for the two parameters, green for AGC and red for VATT. Overlap between the two parameters is therefore indicated by a yellow line. Generally speaking, green lines indicate lower power, diffuse returns whilst red lines indicate higher power more peaked returns. A strong correlation between areas of thin ice and leads and high power peaked returns is observed. Figure 7.2 shows the profile of AGC values and image intensity along the ground track. Some peaks in the AGC over the pack ice region do not correlate with leads visible on the imagery. Given the great sensitivity of the altimeter observations to even small areas of smooth surface, as discussed in chapter 4, it is possible that leads beyond the resolution of the AVHRR are being detected. Lower VATT values over the smaller lead to the east indicate that more peaked returns are occurring than for the wider lead, due to a smoother surface. Another feature is the sharp drop in AGC value corresponding to the fast ice boundary on the Eastern margin. A slight recovery in AGC value is also seen over fast ice nearer the coast.

This comparison clearly demonstrates the sensitivity of the AGC value to changing sea ice roughness. Thin ice is distinguished in the infra-red imagery due to a higher surface temperature, as discussed in section 2.4.1, and in altimeter data due to a smoother surface. In some cases the altimeter may be detecting leads not observed in the imagery.
Plate 7.6 AVHRR Channel 4 image of the Kara Sea 21st February 1986, AGC and VATT values overlayed. Vertical bars for the AGC value are displayed on the red channel and those for the VATT value on the green channel. Where overlap occurs between the two values the bar is displayed as yellow.

Figure 7.2 AVHRR and altimeter profiles for track B on the Kara Sea image. Excellent correlation is observed between high values of AVHRR channel 4, indicating thin areas of ice, and high values of AGC, indicating a smooth surface.
Plate 7.7  AVHRR Channel 4 image of the Chukchi Sea, 10 Feb 1988 01:32, SIGPK and SIGTD parameters overlayed.

Plate 7.8  AVHRR Channel 4 image of the Chukchi Sea, 14th March 1988 16:11, SIGPK and SIGTD parameters overlayed.
From a glaciological point of view, the percentage of thin ice and open water plays an important role in determining the heat flux from the ocean to the atmosphere. From an operational point of view the identification of areas of thin ice and leads is important for ship navigation.

7.3.2 Chukchi sea images

Three mid-winter Chuckchi sea images were available for comparison with Geosat altimetry. Lack of sunlight prevented use of channel 2 data and channel 4 data is therefore presented.

Plate 7.7 shows a channel 4 image acquired over the Chuckchi Sea on the 10th of February 1988. As with the Kara Sea image, an increase in backscattered power and in the rate of fall off of return power (indicated by higher SIGTD values) in the return waveforms is observed over areas of thin ice. This is seen, both for a large area of thin ice north of Wrangels island and also over some narrow leads near smaller islands. Over the pack ice lower back scatter values are observed. Track D passes over an area of large floes, only just distinguishable in the imagery, which are accompanied by low power diffuse returns. Plate 7.8 shows an image of the same area for the 14th of March 1988. Again numerous peaks are seen to correspond with shoreline leads and also over the area north of Wrangels island.

As with the Kara Sea image, both of these images show a clear relationship between altimeter and infra-red observations of smooth ice. In addition we see the first evidence that large floes result in low power diffuse returns as suggested by the discussion in chapter 4. Later, when altimeter elevation data is examined in chapter 9, we shall see that low power diffuse returns within the ice pack correspond with distinct changes in the surface elevation.

7.4 Observations of vast ice floes and fast ice in altimeter and AVHRR data

Plate 7.9 is subscene of the image shown in plate 7.4, showing vast ice floes lying off fast ice on the coast. Cloud cover is observed over the Greenland sea although it is seen to decrease somewhat over the ice pack. The vast floes observed in the image probably originate from high up the the Arctic basin and have been advected down the coast by the East Greenland current. Track B is of particular interest since it passes over two floes with dimensions of several tens of kilometres. Over the larger of the two floes (lying closest to the coast) transected by the altimeter SIGTD values indicate 'ocean' like returns with a peak backscatter
level of around 4-8 dB, significantly lower than that usually observed over the ocean. The smaller floe appears to have moved some distance south-wards, in the direction of the prevailing current, during the 13 hour delay between altimeter pass and image acquisition. A similar fall in backscattered power and return to diffuse waveforms is observed over the fast ice boundary.

Plate 7.10 shows the same image with the leading edge parameters LEWID and LEDIF superimposed. Large excursions in the value of LEWID are observed at the boundaries between large floes and surrounding sea ice. This is a consequence of bright off-nadir returns causing track 'snagging' as the altimeter passes over transitions between vast floes, which result in low backscatter returns, and the bright areas of smaller floes either side. Such excursions are accompanied by large values of LEDIF indicating that the return waveforms have a distorted leading edge. In the centre of the vast floes, however, where low values of LEDIF indicate more well behaved returns, LEWID values are significantly higher than over the surrounding sea ice. Assuming that the backscatter from the ice floe is relatively homogeneous, as is suggested by the time history of SIGPK, the variations in the LEWID parameter represent real observations of ice floe roughness, the first time such measurements have been made by a space-borne sensor.

Figure 7.3 shows image intensity and altimeter parameters along the same track. Delineation of the vast floe boundary is clear both in both altimeter and image intensity profiles. The leading edge width parameter also shows significant modulation over both floes and fast ice, allowing direct estimates of metre scale surface roughness to be made.

To examine this phenomenon more closely one second averages of return waveform profiles were produced and are shown in figure 7.4. Well behaved returns occur over the first vast floe for Time = 18738-18743 and for the second floe for Time = 18750-18756. The differential of the mean waveforms, representing the probability density function of surface height is shown to the right of each mean waveform. For the first vast ice floe the mean waveforms exhibits a significant toe before the main peak. Looking at the differential of the mean waveforms, the distribution of surface heights is certainly reasonable when compared with the distribution of keels shown in figure 2.3 which generally reflect the surface topography but amplified by a factor of 2 or 3. Mean returns from the second floe are more difficult to interpret. Looking at the profiles of LEWID and the surface elevation show that the returns over floe B are less well...
Plate 7.9 Enlargement of plate 7.4 showing SIGPK and SIGTD values over vast ice floes.

Plate 7.10 Enlargement of plate 7.4 showing LEWID and LEDIF values over vast ice floes.
Figure 7.3 AVHRR and altimeter profiles for track B on Plate 7.6. Strong correlations are observed between low AVHRR and low SIGPK and SIGTD values over the two vast ice floes (indicated by arrows) and also over fast ice. The bottom profile shows the elevation relative to the GEM-10B geoid. Although it is hard to say whether a significant increase in elevation occurs over the vast floes, the plot does demonstrate the low noise level in height measurements.
Figure 7.4a Mean waveform profiles over floe A (fig. 7.3).

Figure 7.4b Mean waveform profiles over floe B (fig. 7.3)

Figure 7.4 Mean waveform profiles over vast ice floes. Each plot represents the mean of 10 waveforms as they appear in the telemetry. Mean profiles of floe A show a significant toe on the leading edge which may correspond to ridges. Mean profiles of floe B are significantly different in shape possibly due to a different height distribution or to penetration of a thick snow layer.
behaved than for floe A and may be caused in part by the slope observed in the elevation profile shown in figure 7.3. Nevertheless the mean return waveforms show a shape characteristic of penetration of the surface due to a thick covering of snow [Ridley and Partington, 1988].

The elevation profile shown in figure 7.3 was produced by subtracting a second order polynomial derived by comparison with elevation values which appear on the corresponding GDR. Although it is hard to say that a significant difference in surface elevation occurs over the floes, as compared with the surrounding water, the low level of noise over the ice floes demonstrates that elevation measurements with a precision approaching that attained over the ocean should be possible.

The measurement of ice freeboard is potentially very important in providing the estimates of ice thickness needed for model validation and other applications discussed in chapter 1. Although in this case, measurements have only been made over a very small percentage of the ice, the coverage afforded by ERS-1 will reach far into the Arctic basin where a much larger population of such ice floes exists. Such measurements will depend on the accurate determination of ocean surface height either side of such floes but this should be achievable over a long time period. The determination of surface elevation areas is further addressed in chapter 9.

As with the Kara Sea image, then altimeter is seen to accurately delineate the fast ice boundary. The slight recovery in backscattered power further into the fast ice may reflect the boundary between fast ice which has survived the entire winter and that which has broken up at some point.

7.5 Comparison of Geosat altimetry with airborne SAR data collected during MIZEX-87

On the 13th of April 1987 during the Marginal Ice Zone EXperiment (MIZEX-87) an underflight of several Geosat altimeter tracks was made by an aircraft carrying the Interra Star 2 system. An initial comparison of AGC and VATT values with the SAR data was carried out by Fetterrer et al. [1988]. In this section we take that analysis a stage further by looking at waveform
parameters derived from the Geosat SDR and WDR data.

Figure 7.5 shows the SAR image acquired over some 450 km of ice with a swath width of the SAR is 63 km. The two thin white lines running the length of the image show the approximate area covered by the altimeter assuming a footprint of radius ~4.5 km during about 60 seconds. Geo-location of the SAR image was carried out using points on the Greenland coast in comparison with large scale maps and the error is estimated at less than 1 km. The across track error in the altimeter position is within 100 m but the along track error may be one or two kilometres. The time difference between satellite and SAR aircraft underflight is less than an hour for the entire segment. Movements of the ice pack during this time should be small and certainly less than 1 km.

7.5.1 Airborne SAR observations of sea ice off East Greenland

As with the altimeter the SAR instrument is sensitive to roughness variations on scales comparable to the wavelength at which the instrument operates (~6 cm). Since the SAR observes the sea ice at angles far from nadir (60-85°) only very rough surfaces will appear bright in the image. This simple principle helps somewhat in interpretation of the image.

The first 100 km or so shows a fairly uniform bright appearance with individual ice floes unresolved which may be interpreted as the marginal ice zone. In the zone approximately 100-200 km into the edge some large floes are resolved interspersed with bright areas of smaller floes or brash. Beyond 200 km the ice floes become more compact and show a dark uniform appearance suggesting a fairly smooth (on the metre scale) roughness. At around 300 km a vast old multi-year ice floes lies centred on the altimeter track. Its bright appearance and drainage channels visible on the surface suggest that this is a very old floe probably originating in the central Arctic. Between this point and the fast ice boundary, observed as a bright line crossing the image, many more bright floes are observed showing that this is a very mixed ice regime. Within the fast ice the first 20 or so kilometres appear quite bright and are followed by a transition to a darker smoother surface. A likely explanation is that the outer zone represents fast ice which has been broken up in winter storms and later refrozen whilst the inner zone has survived the whole winter.
Figure 7.5  Geosat altimeter data in comparison with airborne SAR data collected during MIZEX-87. Two thin white lines show the 9km swath of the Geosat altimeter range window footprint. Changing altimeter parameters correspond to different sea ice zones observed in the airborne SAR data. In addition some smaller features, such as leads, vast ice floes and the fast ice boundary are observed in both.
7.5.2 Comparison of altimeter backscatter with airborne SAR observations off East Greenland

The upper plot of figure 7.5 shows the waveform backscatter parameters, SIGPK, SIGTD and SIGTR. The ice edge transition, which is unfortunately missed by the SAR image, is shown by a rapid rise in SIGPK from the open ocean value of 9dB to a level around 25-30dB. In the first 100km (time ~ 24340-24344), or so, corresponding to the MIZ, SIGPK values lie in the range 25-35 dB. SIGTD values are also fairly steady with a value around 5dB. SIGTR values are low (1-2dB) throughout this segment and suggests that waveforms here closely follow the Physical optics stationary phase approximation (Gaussian slope distribution) given in section 4.5.3. This contrast with the transect over East Greenland by Seasat which is discussed in section 6.2, where SIGTR values are seen to rise to values between 2-4 dB further into the pack. The fact that the highest backscatter values occur in the region of highest intensity in the SAR image certainly suggests two distinct components to the surface roughness. Inter-floe distance in this region is probably only a few metres and as discussed in section 4.7.1.2, the damping of wind waves by the ice floes and the limited fetch over which such waves can be regenerated is likely to lead to a smooth water surface with only the long period swell penetrating.

Significant features are again observed in both imagery and altimeter data. As the altimeter passes over a lead (a) of significant size, at around 100km (time ~ 24334s) into the edge, a dip in the backscatter is observed showing that the fetch of one or two kilometers has allowed regeneration of wind waves. As the altimeter passes onto the larger smooth floes the probability of smooth water areas lying near the nadir point decrease and lower more variable values of SIGPK are observed. A further decrease in backscatter is observed in the last region (200-300km; time ~24360-24376s) corresponding to the increase in average floe roughness. Flat steady backscatter values of around 15dB are observed over several data points covering a large first year floe (b) which lies almost centred on the altimeter track. Another significant dip in SIGPK, to below 10dB, is seen over the rougher multi-year ice floe (c). A sharp fall in SIGPK is seen as the altimeter passes onto the fast ice (d) with SIGTD and SIGTR indicating a return to diffuse, 'ocean like', returns. A slight recovery of around 4dB is also seen as the altimeter crosses onto the smoother fast ice further into the fjord.

The lower plot on figure 7.5 shows two leading edge parameters LEWID and LEDIF. Over the marginal ice zone both parameters continue at steady low values indicating steep leading
edges which usually accompany peaked returns. In the area of larger first year floes some excursions in the LEWID parameter are observed as off-nadir bright areas are observed over larger floes. In the mixed multi- and first year ice regime large excursions are observed. Over the large first year ice floe (b) sharp spikes delineate the transitions onto and off the ice floe caused by bright targets off-nadir. The source of high LEWID values in the middle of the floe is less clear. A region of sustained high LEWID values is also observed over the large multi-year floe (c). Although, these are in part due to off-nadir bright areas they also reflect the degree of surface roughness of the ice floe. A further large excursion is observed at the fast ice (d) transition followed by steady values around 2m providing direct measurements of the roughness of the fast ice in the fjord.

7.5.3 Discussion of SAR/Altimeter comparison

In summary the altimeter seems to provide discrimination of four zones within the sea ice off East Greenland:

(i) The marginal ice zone
(ii) Large smooth first-year floes
(iii) Mixed first- and multi-year ice
(iv) Fast ice

In addition the altimeter is sensitive to leads within the MIZ, first and multi-year ice floes covering at least the range window footprint and different zones within the fast ice. This comparison adds further weight, therefore, to the nature of altimeter signals over different ice zones observed in the AVHRR comparison.

The widely different sampling scales of the two instruments makes a more quantitative comparison difficult. In addition the uncertainty in geo-location (~1-2km) limits the scope for comparing the response of the altimeter to small (<100m) smooth areas of ice observed in the SAR image.
7.6 Conclusions

Although the datasets shown here have been small, significant correlations have been observed between altimeter and AVHRR/SAR observations of sea ice. Further work is obviously required to examine regional and seasonal differences which may effect the altimeter response to different type of sea ice. The technique of combining image and altimeter data clearly allows more information to be extracted than would be possible with either data set in isolation. The altimeter and ATSR instruments on board ERS-1 should provide co-incident coverage and will allow significant advances to be made.

To summarise the results of this chapter, the altimeter has shown clear delineation of several features, in particular leads and areas of thin ice, vast ice floes and the fast ice boundary. Since the scale of these features is frequently beyond the resolution of Passive Microwave instruments, radar altimeters may provide the best means for the global mapping of such features until such a time that global data from space-borne SAR systems becomes available.

As far as overall sea ice mapping is concerned the Geosat/MIZEX SAR comparison shows a correspondence between changes in the altimeter signal and different zones within the ice. As with the results from the analysis of the Seasat Polar Reference Data Set, the highest backscatter values are normally observed near the sea ice edge corresponding to the Marginal Ice Zone (See section 2.1).
8.0 Introduction

In this chapter we present results from analysis of 26 months of data from the Geosat Geophysical Data Record. Although no waveform profiles are included in this data set sufficient parameters exist to permit coarse location of the sea ice/ocean boundary. By mapping the sea ice boundary for 48 seventeen day repeat cycles, seasonal and inter-annual variations in the total area enclosed by the sea ice boundary over a period of more than two years can be produced. The data set employed in this analysis covers the period from November 1986 to February 1989. The results are compared with the total Southern hemisphere sea ice extent as mapped by passive microwave observations. Agreement between passive microwave and altimeter derived total sea ice area during the 'freeze up' period is excellent. During the 'melt period', however, altimeter total ice area is consistently higher than that mapped by the passive microwave instrument.

8.1 Geosat ocean GDR global coverage

The format of the Geosat GDR is described in section 5.1.2. The study employs the Geosat 'ocean' GDR only. Unfortunately much of the data over sea ice is edited from the 'ocean' GDR and placed into the 'land/ice' GDR according to criteria specified in Cheney et al. [1987]. The filter applied is designed to eliminate poor data points over the ocean and uses checks on various tracker parameters which generally eliminate highly peaked returns. Figure 8.1 shows the coverage of the ocean GDR during cycle 27 of the Exact Repeat Mission (ERM). In some land areas, however, such as large inland lakes or deserts, the presence of 'ocean like' data (as observed by Rapley et al. [1987]) is clearly demonstrated.

Figure 8.2 shows GDR coverage of the Southern hemisphere during the first ERM cycle, data editing clearly poses a problem when using this data for sea ice mapping. Coverage up to the sea ice boundary is, however, normally achieved. Large areas of data also remain over ice shelves such as the Larsen and off the Eastern coast of the continent.
Figure 8.1 Geosat 'Ocean' GDR global coverage for ERM cycle 29.

Figure 8.2 Geosat 'Ocean' GDR Southern Ocean coverage for ERM cycle 1.
8.2 Data averaging prior to classification

As discussed in chapter 3 the presence of sea ice causes disruption of range and significant wave height measurements and is also normally accompanied by an increase in AGC values. The discussion of sea ice edge detection in chapter 7 showed that further work is required to fully understand the altimeter response to compact and diffuse ice boundaries. It is, however, reasonable to assume that, on average, significant disruption to the parameters recorded in the geosat GDR will occur within 10 or 20 km of the ice edge. This accuracy was considered sufficient to carry out a pilot study of the potential of satellite radar altimeter for mapping the seasonal and inter-annual variations in sea ice extent. It should be stressed that this study is viewed as a prelude to a more detailed study which should employ a full waveform data set.

Since the volume of data involved is quite large (approaching 1 GByte globally) it was decided to first average GDR parameters into geographical bins. The choice in the size of bin requires a trade off between the maximising the resolution of the averaged data set from which the ice edge was to be determined and ensuring that sufficient data points exist within each bin, taking into account data editing, to give a reasonable statistical measure of changes in the altimeter signal. Since the sea ice boundary, on average, follows a line of constant latitude, it is preferable for bins to be smaller in the latitude direction and larger in the longitude direction. Ensuring that bins are large enough to contain several data points is a difficult problem, given that the data editing is not necessarily systematic. The final size was chosen, through some degree of trial and error, to be 0.4° latitude x 2° longitude. The longitude dimension ensures that at least two orbit tracks pass through each bin and the latitude sample is chosen to be similar in resolution to passive microwave instruments. For each bin the following GDR parameters were averaged.

(i) SDH - One second standard deviation of surface elevation
(ii) AGC - Automatic Gain Control value
(iii) SWH - Significant Wave Height value
Figure 8.3. Average GDR parameter values for ERM cycle 1. All three parameters show a significant change close to where the ice boundary is expected. Average values for the 1s standard deviation show the most constant values over the open ocean since, unlike AGC and SWH, it does not vary due to changing sea conditions.
8.3 Selection of suitable thresholds for sea ice detection

Figure 8.3 shows plots of the averaged values of three parameters; SDH, AGC and SWH from ERM cycle 1. The AGC value, which was used by Rapley [1984] to indicate the presence of sea ice, shows higher values in regions where sea ice cover is expected. Significant variance in the AGC value is observed over the open ocean making selection of a universal threshold to indicate the presence of sea ice difficult. The SWH value shows a sharper transition from normal to extreme values resulting from mistracking over sea ice. The ice boundary as mapped by SWH values is less well behaved than that mapped by the AGC value, indicating that changes in waveform shape which cause extreme tracking problems may not occur until some distance into the ice edge. Again normal variations in the value of SWH are observed over the ocean making choice of a threshold applicable to all ERM cycles more difficult. The SDH value shows an ice boundary similar to that indicated by the AGC and SWH but response over the ocean is essentially flat making choice of a universal threshold more practical.

![Histograms of AGC, SWH and SDH for binned data from ERM cycle 1.](image)

Figure 8.4 shows a histogram of SDH, AGC and SWH values of individual geographical bins. The profile of SDH shows a very sharp peak around 0.03 with a much lower but more even distribution from around 0.06 up to 1.00. AGC values show a much broader distribution, a
sharp fall off around 30dB probably lies close to the cutoff between ocean and sea ice data although some overlap is likely. SWH values also show a fairly wide variation around a peak of 10m. Again sea ice and ocean signals probably overlap considerably.

From the plots and histograms of the three parameters, shown in figures 8.3 and 8.4, SDH seems to provide the best discrimination of ocean and sea ice although SWH and AGC values can provide secondary indicators of the presence of sea ice. The threshold on SDH should also be set high enough to avoid classifying as sea ice slight glitches in elevation values over the ocean ocean areas caused by slicks, low wind speed areas or rain cells [Laxon and Rapley, 1987]. A geographical data bin is therefore classified as ocean if one of the following thresholds is exceeded:

\[
\begin{align*}
\text{SDH} & > 0.01 \text{ m} \\
\text{SWH} & > 20 \text{m} \\
\text{AGC} & > 35 \text{ dB} \\
\end{align*}
\]

8.1

The optimisation of the thresholds could be carried out by minimising the difference between the ice boundary derived using this technique and that determined from some other sensor over a large number of crossings. The above thresholds were therefore selected as a first approximation.

8.4 Classification of data bins

To compute the total area enclosed by the sea ice boundary first requires each bin to be classified as one of the following:

(i) Ocean
(ii) Land
(iii) Sea ice
(iv) Unknown

Discrimination between ocean and land data is required since data over land will exhibit similar disruption of altimeter parameters as occurs over sea ice. In addition a certain fraction of the
data must remain unknown where ocean exists beyond the latitudinal limit of 72.08°. Where data editing has resulted in a bin with no data points the bin is classified as being identical to that with the same longitude value and a latitude value one increment (0.4°) Northward. Figure 8.5 gives a flow diagram of the algorithm used to classify each data bin.

![Flowchart showing algorithm used to classify data bins of geographically averaged Geosat GDR data.](image)

**Figure 8.5** Flowchart showing algorithm used to classify data bins of geographically averaged Geosat GDR data.
Figure 8.6 shows the classification of data bins for ERM cycle 8. Unknown areas are seen mainly around the Ross sea. In addition around ten data bins seem be scattered outside the main ice boundary. These are most likely caused by anomalous events over the ocean or inaccuracies in the land mask used.

To compute total area of data bins containing sea ice each latitude longitude bin are approximated by a trapezoid and the area of each is computed using:

$$\text{Area} = \text{DLat} \times \frac{(\text{DLong}_1 + \text{DLong}_2)}{2}$$

Where $\text{DLat}$ is the length (in km) of the trapezoid in the North-South direction and $\text{DLong}_1$ and $\text{DLong}_2$ are the lengths of the trapezoid in an East-West direction for the South and North edges. $\text{DLat}$ is computed using:

$$\text{DLat} = \Delta \text{Lat} \times 110 \text{ km}$$

$\text{DLong}_1$ and $\text{DLong}_2$ are computed using:

$$\text{DLong}_1 = 110. \times \Delta \text{Long} \times \cos(\text{Lat})$$
$$\text{DLong}_2 = 110. \times \Delta \text{Long} \times \cos(\text{Lat} + \Delta \text{Lat})$$

The total area enclosed by the sea ice boundary for each ERM is then computed using:

$$\text{TOTAL AREA} = \frac{\text{TOTAL AREA OF SEA ICE BINS} + \text{TOTAL AREA OF UNKNOWN BINS}}{2}$$
Figure 8.6 Classification of geographical bins for ERM cycle 8. The classification of areas is as given in the key. Areas mapped as unknown make up a fairly (<5%) small percentage of the total. Some geographical bins over the ocean have been misclassified due to inadequacies in the land mask used, and the presence of anomalous events.
The error assigned to the TOTAL AREA is computed using:

\[
\text{ERROR} = \frac{\text{TOTAL AREA OF UNKNOWN BINS}}{2} + 110^2 \times 360 \times 0.2 \times \cos(65^\circ)
\]

the additional term for the error assumes an average latitudinal error of 0.2° and a mean latitude for the sea ice boundary of 65°.

8.5 Comparison of total Antarctic sea ice extent measures by satellite altimetry and by passive microwave observations

Figure 8.7 shows the total sea ice extent derived using the method described previously compared with that published by Gloerson and Campbell [1988] using passive microwave (SMMR) imagery. Overlap between the two plots occurs from the beginning of the ERM mission to July 1987. During the Austral summer 86/87 the total area enclosed by the sea ice boundary as mapped by the altimeter is seen to lag behind that mapped by the SMMR. The discrepancy is typically 1-2 million square kilometres. If we assume that the mean latitude of the sea ice boundary, near minimum, is 70°S this corresponds to a discrepancy in distance of 75-150km. The minimum area reached by the altimeter is also slightly greater than that reached by the SMMR. Shortly after the minimum sea ice extent is reached, however, the match between the two areas becomes very close.

There are two possible geophysical reasons for the difference in altimeter and SMMR total sea ice extents during the melt period:

(i) The altimeter is detecting changes in the ocean surface prior to the sea ice boundary due to meltwater which may cause damping of small wind waves.

(ii) The SMMR does not detect low sea ice concentrations occurring near the ice edge due to melting of the surface of ice floes.
Figure 8.7. Total Antarctic sea ice extent mapped using Geosat GDR data (November 1986 to February 1989) compared with that mapped using SMMR data (January 1985 to July 1987, adapted from Gloerson and Campbell, [1988]). Agreement during the freezing phase (March - July 1987) is remarkably good, but significant differences are observed during the preceeding melt period.
Although option (i) is a possibility it seems unlikely that cold meltwater could extend out to 100 km beyond the ice edge before heating up sufficiently to avoid the surface layer effect. It is well known that due to uncertainties in ocean surface roughness that the minimum ice concentration that can be reliably detected using SMMR is 15% [Carsey, 1982]. This problem becomes worse during periods of melting when liquid water on ice floes reduces the brightness temperature contrast between open ocean and sea ice considerably [Carsey, 1985]. Option (ii) therefore seems the more likely but direct comparison of altimeter and passive microwave derived sea ice extents would help considerably.

8.6 Conclusions

The seasonal variation in total Antarctic sea ice extent has been mapped over a period of more than two years using data from the Geosat Geophysical Data Record. Although resolution is compromised by missing data points and coarser sampling than may be possible from the 10/s Sensor Data Record it is sufficient for a hemisphere wide comparison with passive microwave results. Comparison of the measurements of total Antarctic sea ice extent using the altimeter and SMMR show close agreement during the freeze up period but the altimeter derived total ice area is generally larger than that derived during the melt period. A direct geographical comparison of the ice extent mapped by the two sensors is required to more fully understand the differences observed.

This study is intended only as a demonstration of the altimeter’s capability for global sea ice extent mapping. It shows that satellite radar altimeters can provide a synoptic, independent measure of the global sea ice extent, previously only possible using passive microwave instruments. As discussed in section 1.2.6, sea ice extent may provide a sensitive indicator of changes in global temperature, and is therefore a parameter of great importance.
9.0 Introduction

So far we have looked at the ways in which the shape and strength of the return signal from a space-borne altimeter responds to changing sea ice conditions. In this chapter we concentrate on the elevation information from radar altimeters over sea ice. This work aims to explore the potential of radar altimeters in mapping the marine geoid in of sea ice areas, and in providing measurements of ice freeboard. Following the sequence of analysis carried out in chapter 6, we shall first look at elevation data from the three day repeat cycle, and then look at data from the Polar Reference Data Set. In the last section we show that radar altimeters can also provide measurements of the changing freeboard of giant tabular icebergs.

9.1 Measurements of mean sea surface elevation by satellite radar altimetry

Over the open ocean the Seasat and Geosat altimeters can provide measurements of the delay time between the transmission and receipt of a radar pulse with a precision, when averaged over one second, equivalent to a few cm of range. However, the accuracy of the range measurement depends on knowledge of atmospheric conditions which affect the propagation time of the radar pulse. The true range, \( h \), to the surface is given by:

\[
h = h_{\text{measured}} - \Delta h_{\text{ion}} - \Delta h_{\text{trop}}
\]

Where \( \Delta h_{\text{ion}} \) and \( \Delta h_{\text{trop}} \) are the corrections for the changing radar path length caused by the ionosphere and troposphere. Once the corrected height measurement is obtained the elevation of the surface above the reference ellipsoid, \( H \), is calculated using:

\[
H = h - H_{\text{sat}}
\]

where \( H_{\text{sat}} \) is the altitude of the satellite above the reference ellipsoid.
The accuracy to which $H_{sat}$ is determined is limited by the accuracy to which the Earths' gravitational field is known. Values of $H_{sat}$ included in the Seasat Geophysical Data Record show a RMS radial error of about 1.5m [Marsh and Martin, 1982]. Since many oceanographic and geophysical features have signatures of similar or smaller magnitude, it is important to reduce this error. Fortunately the largest components of orbit errors vary over wavelengths of several thousand km whilst signatures due to ocean bathymetry and oceanographic features occur over much shorter distances. A common method of reducing orbit errors is to apply a bias and tilt to each orbit arc in a small region, in such a way that the total differences in measured surface elevation at the points where different orbits intersect (a technique known as cross over analysis) is reduced [Zwally et al., 1983c]. In the analysis presented here, the technique of bias and tilt removal is applied to different orbits covering the same ground track during the three day repeat cycle. In the analysis of elevation data from the Polar Reference Data Set a new orbit determined by Marsh et al. [1988] is used with an RMS error of around 50 cm. Since the errors in altimeter elevation measurements over sea ice are of a comparable or greater magnitude it was decided that little improvement would be gained from applying cross over analysis to this data set.

The elevation of the sea surface above the reference ellipsoid depends on the height of the marine geoid, $H_{geoid}$ and on the dynamic sea surface topography, $H_{dynamic}$, thus:

$$H = H_{dynamic} + H_{geoid}$$  \hspace{1cm} 9.3

The height of the marine geoid varies by as much as ±100m around the entire globe and depends on the gravitational effects of varying bathymetry and more deep seated mass anomalies.

The components of dynamic sea surface topography can be further broken down into the following components [Calman, 1987]:

$$H_{dynamic} = H_{mc} + H_{meso} + H_{tide} + H_{wave} + H_{bar} + \Delta H$$  \hspace{1cm} 9.4

where:
\[ H_{mc} = \text{height due to mean currents} \]
\[ H_{meso} = \text{height due to transient currents and mesoscale features, such as eddies.} \]
\[ H_{tide} = \text{height due to tidal effects} \]
\[ H_{wave} = \text{height due to changing surface waves} \]
\[ H_{bar} = \text{height due to changing atmospheric pressure} \]
\[ \Delta H = \text{height due to any other time variant contributions} \]

with the exception of \( H_{mc} \), all of these components will be time variant. The amplitude of \( H_{mc} \), \( H_{meso} \) may be as much as a metre or so, for a current such as the Gulf Stream and its associated eddies. Tidal differences are normally small over the open ocean but can be several metres near coasts. The contributions of the other components is normally a few centimetres.

Unless otherwise stated all elevation data presented in this chapter has either the PGS-S3 (Seasat data) or GEM-10B (Geosat data) geoid removed.

9.2 Seasat elevation profiles obtained during the three day repeat cycle

As discussed in chapter 3, elevation values from the Seasat and Geosat altimeters contain significant errors caused by deviation of the return waveform shape from the observed over the open ocean. To reduce mistracking errors it is necessary to add a correction, determined from the offset of the waveform leading edge and the tracking point, and also to eliminate 'height glitch' events (see chapter 5). In the first part of the analysis of the three day repeat cycle, the improvement in the accuracy of surface elevation measurements using these techniques is assessed. In the second part the correlation between height deviations and changes in peak backscatter is examined to see if measurements to the ocean surface and to the top of ice floes can be distinguished.

9.2.1 Assessment of altimeter retracking over sea ice

Ideally validation of geoid retrieval by retracking of altimeter data should be carried out by comparing profiles from exact repeat tracks during times of minimum and maximum sea ice extent. Over a distance of one or two kilometers the variation in the sea surface elevation as a result of the geoid is negligible [Brown et al., 1983]. Because the Seasat mission covers
only a three month period with a change in orbit pattern after two months this technique is difficult to apply to data from the Seasat altimeter. The approach adopted here is therefore to determine the deviation of individual elevation profiles, collected during the three day repeat cycle, with their mean.

Any residual between the elevation determined at a particular point on the track can be the result of several effects. The largest of these will usually be due to the satellite orbit and the effect of long wavelength tides. These can be eliminated by performing a least squares fit to the residual between the two sets of elevation measurements. This least squares line is then subtracted from the residual profile to remove the effects of the tilt and bias on each orbit [Thompson et al., 1983]. Any discrepancy remaining in the elevation residual is due to either real effects or due to errors in the retracking algorithm.

Figure 9.1a shows raw and retracked elevation profiles from the inward transect of the three day repeat orbits over the same track that was discussed in chapter 6. Individual profiles have been offset by 20m to aid clarity. The mean profile is obtained by averaging elevation values in one second bins and subsequently applying a 3 point (0.25,0.5,0.25) smoothing filter. Figure 9.1b shows the raw and retracked profiles after removal of the mean elevation and a tilt and bias, determined using a least squares fit line to the retracked elevations, has been removed. The reduction in noise on the corrected, compared with the uncorrected, elevation measurements is immediately apparent. The characteristic tracker oscillation which results in a positive deviation of elevation measurements, has been virtually eliminated. Negative deviations, caused by tracker snagging still remain but are fairly infrequent.

Figure 9.1c shows histograms of the elevation residuals for the corrected and uncorrected profiles. The noise and bias on the elevation measurements has been significantly reduced. Overall RMS deviations, indicated by the figures shown next to each histogram have been reduced from around 1.1 metres to 0.45 metres in all the examples shown. Although this does not include any constant biases, which may have been included, this result shows that the detection of small scale geoid features, such as sea mounts and ocean trenches, will be significantly improved by this technique.
Figure 9.1a  Raw (thin) and retracked (thick) elevation profiles and mean from 3 day repeat cycle data. Elevation profiles for each of the repeat cycle have been offset to aid clarity. Units in metres.

Figure 9.1b  Residual elevation profiles after subtracting the mean profile shown in figure 9.1a. Excursions in the raw (thin) elevation profile due to tracker oscillation (similar to those seen in figure 5.1) can be clearly seen. Units in metres.
Figure 9.1c Histograms of raw (thin line) and retracked (thick line) residual elevation values for the inward pass of the three day repeat cycle track. Numbers next to each histogram refer to the corresponding total RMS deviation.
9.2.2 Elevation profiles in comparison to peak backscatter

Fig 9.2a shows a series of corrected elevation profiles from the outward pass of the 170°E data set. Fig 9.2b shows the elevation residuals, with the mean value removed, along with the peak backscatter value plotted on a scale from 0 to 40 dB. Fig 9.2c show scatter plots of peak backscatter versus residual elevation for each of the 4 passes.

A tight cluster of points is seen in each pass, at a backscatter level of around 10 dB, corresponding to returns over the ocean. At higher backscatter levels the spread of elevation measurements becomes larger due to errors which still remain after retracking. Two sections of constant backscatter are observed within the ice pack, in the passes for the 18/9 and 21/9. The 18/9 area shows a lower backscatter level and higher elevation than for the open ocean, as is revealed more clearly in the scatter plot. Another section of ocean like returns is seen in the 21/9 pass this time with a level of peak backscatter and elevation similar to that seen over the ocean.

Given the evidence presented in previous chapters, suggesting that low backscatter areas are associated with vast floes or areas of compact ice, the elevation difference observed in the 18/9 pass seems likely to represent a measurement of ice freeboard. The constant backscatter area in the 21/9 is may either be open water, or may represent thinner ice.
Figure 9.2a  Retracked elevation profiles for the outward pass of the three day repeat cycle track. Units in metres.

Figure 9.2b  Residual elevation profiles after subtracting the mean profile shown in figure 9.2a. Units in metres. The thick line shows the residual elevation whilst the thin line shows the peak backscatter value for each pass on a scale from 0 to 40 dB.
Figure 9.2c  Scatter plots of peak backscatter versus residual elevation. The 18/9 track shows a significant positive residual elevation over an area of low power returns, characteristic of vast floe or compacted ice.
9.3 Mean surface elevation over sea ice in the southern ocean

In this section we analyse elevation measurements carried out by the Seasat altimeter during the 17 day period covered by the PRDS. A database constructed from the Geophysical Data Record allows atmospheric corrections to be applied and removal of the GEM 10-B geoid if desired. A refined orbit with an estimated RMS error of 50cm [Marsh et al., 1988] is employed in the analysis.

9.3.1 Raw and Retracked elevation noise

Figures 9.3a and 9.3b show the one second standard deviation of surface elevation (PGS-S3 geoid [Marsh and Martin, 1982] removed) in the Antarctic before and after retracking and editing of data which failed one or more of the tests specified in section 5.3. Comparing the two plots a significant improvement in the one second standard deviation is observed. For the retracked data, noise values are generally lower in the periphery than elsewhere. Further into the ice pack the distance between areas of new ice and open water becomes greater making tracker 'snagging' (section 3.6.3) events more likely. Again in the two regions of low backscatter, in the Belinghausen sea and off the Oates coast, levels of noise similar to that normally observed over the ocean are observed.

Figure 9.4a shows the one second standard deviation of raw elevation values for the Arctic. Comparing with figure 9.4b which shows the area which Anderson et al. [1988] suggests will be unaffected by sea ice, the disruption of altimeter elevation values is seen to extend much further south than he suggests. Although the PRDS does not correspond exactly to the minimum extent in the Arctic, it is fair to say that Andersons' map of areas that will be ice free at some time is probably optimistic.
Figure 9.3a One second standard deviation of raw elevation measurements (metres) for the Antarctic.

Figure 9.3b One second standard deviation of retracked elevation measurements (metres) for the Antarctic.

Figure 9.4a One second standard deviation of raw elevation measurements (metres) for the Arctic. Disruption of altimeter returns occurs over a wider area (in particular in the Beaufort Sea) than is suggested by figure 9.4b.

Figure 9.4b Extent of disruption to radar altimeter data from ERS-1 over the Arctic ocean assumed by Anderson [1988].
9.3.2 Southern ocean mean sea surface elevation derived from Seasat

Figure 9.5a shows Southern ocean bathymetry from Gordon and Molinelli [1982]. Whilst figure 9.5b shows the mean sea surface elevation derived from the retracked Seasat dataset with respect to the reference ellipsoid. To enhance shorter wavelength features the plot of elevation has been smoothed and differentiated from left to right. The quality of the image is compromised by the presence of data dropouts which appear as bright squares with dark shadows to the right. Nevertheless some features clearly stand out. Sections of the South Sandwich trench clearly stand out to the North and East of the Antarctic peninsular. Features mapped below the sea ice include the continental shelf in the Belinghausen Sea and a significant ridge, running parallel to the Larsen ice shelf. The Larsen ridge is not mapped in the bathymetry shown in figure 9.5a, and may represent a previously undiscovered geophysical feature of considerable importance [P. Berry, personal communication].

These plots demonstrate that, by employing retracking techniques, satellite altimeter data that was previously excluded by some researchers (e.g. Marsh and Martin [1982]) can be used to map significant features of the marine geoid where sea ice cover exists. Although observations of the geoid by the Geosat altimeter carried out during the Antarctic ice minimum will overcome the problem of sea ice contamination for most of the Southern ocean, some areas will remain ice covered throughout the year. A more important area, in which this technique may be used, is the Arctic basin where many geophysically important areas may lie under permanent ice cover.
Figure 9.5a Southern Ocean bathymetry from Gordon and Molinelli [1982]. Features identified here and in figure 9.5b are: (A) Belinghausen sea continental shelf, (B) Trench north of the South Orkney islands, (C) Shelf/Ridge off the Larsen ice shelf, (D) South Sandwich trench, (E) Maud rise, (F) Oates coast continental shelf.

Figure 9.5b Mean sea surface topography mapped using retracked Seasat data. The image intensity shows the smoothed differential of the observed elevation (no geoid removed) performed from left to right. Moving from left to right positive elevation changes are brighter than average and negative elevation changes are darker than average. The image shows several bathymetric features identified in figure 9.5a.
9.3.3 Elevation profiles off the Oates coast

Analysis of the 17 day repeat dataset in chapter 6 showed that a large area of 'ocean' like returns, associated in chapter 7 with reflections from vast multi-year floes, existed off the coast of Oates Land. A more detailed examination of elevation data is shown in figure 9.6. Each data point is offset vertically from the position of the ground track according to the value of the elevation at that position. To aid clarity, descending passes only are plotted.

At lower latitudes the elevations profiles are observed to be quite flat, if a bit noisy. Square dips appearing in the profiles are caused by sequences of data that have been excluded by the waveform quality control algorithm (section 5.3). As the tracks run in towards the coast a gradual rise of one or two metres is observed corresponding to the edge of the continental shelf as seen.

Of particular interest are orbits 575 and 618 which show steps in the elevation profiles over the continental shelf. Figure 9.7 shows the individual elevation profiles along with various waveform parameters. Data over these two features consists of diffuse low power returns which were shown in section 7.4 to correspond with observations of vast multi-year ice floes. No instrumental effect can account for these changes and the rises occur too rapidly to be caused by bathymetry, since even sharp changes in bathymetry results in a signal smoothed over tens of km on the ocean surface. The most likely explanation that the elevation differences observed represent real changes in the freeboard, and hence thickness, of ice floes in that region.
Figure 9.6 Retracked elevation profiles (GELEVR) off Oates land. Data points are vertically offset from the satellite ground track by a distance proportional to the retracked elevation (Geoid subtracted). An approximate vertical scale is shown for orbit 618. Elevation profiles consistently show a rise as they approach the coast, presumably do to the effects of the continental shelf. Elevation profiles on some orbits (e.g. orbits 575 and 675) show step height changes, several metres above the underlying trend. Sharp features such as this are difficult to attribute to changing bathymetry and the changes observed may be due to radar returns from the top of thick ice floes.
Figure 9.7a Elevation profile and altimeter parameters for Orbit 575.

Figure 9.7b Elevation profiles and altimeter parameters for orbit 618.

Figure 9.7 Step changes in the elevation are accompanied by low backscatter, 'ocean' like returns which have been associated with altimeter returns of vast floes (section 7.4).
9.4 Iceberg freeboard measurements

McIntyre and Cudlip [1986] present observations of a giant tabular iceberg using data from the Seasat altimeter. They showed that the return from the iceberg surface was similar to that over the ocean and obtained elevation profiles for almost the entire length. In this analysis we look at elevation measurements over two giant tabular icebergs that were present in the Southern Ocean during the time of the Geosat mission. As with the analysis of ice extent, presented in the previous chapter, the Geosat GDR data are used since they provide a much larger data set for possible iceberg crossings than would otherwise be the case.

Between February and March 1986 a large section of the Larsen ice shelf broke off forming a giant (95x95km) iceberg which was subsequently responsible for the calving of a smaller section of the ice shelf just to the north [Jacobs et al., 1986; Jacobs and Barnett, 1987]. The point at which the calving took place corresponds with an area of ice shelf ridging, observed previously by Ridley et al. [1989] using data from the Seasat altimeter. At some points the large iceberg split into two smaller icebergs, hereafter referred to as Larsen A and Larsen B. Figure 9.8 shows an image of the two icebergs acquired around January 1988. The two icebergs moved slowly North over a period of many months with their position regularly monitored by infra-red and passive microwave observations and shown on the US Navy/NOAA ice charts.

9.4.1 Iceberg identification

The most practical method for identifying large icebergs in satellite altimeter data is to look at individual elevation profiles. The freeboard of large tabular icebergs is typically several tens of metres, significantly greater than the overall deviation of the mean sea surface from the geoid. Figure 9.9a shows the position of the Larsen A (A-20A) iceberg identified on the NAVY/NOAA Joint Ice Centre for the 11th of February 1988. Figure 9.9b shows elevation profiles for the 27th ERM cycle from Geosat for an area Northeast of the Antarctic peninsular. Elevation values are not corrected for tracking errors and have had the Atmospheric correction, tide and Gem10-B geoid removed. A step is observed in the elevation values at a position corresponding closely to the position of the Larsen A iceberg as mapped in the ice chart.
Figure 9.8  Larsen A and B observed in AVHRR imagery in January 1987 [Jacobs and Barnett, 1987].

Figure 9.9a  Navy/NOAA ice chart 11th Feb 1988 showing the position of the Larsen A iceberg (A-20A).

Figure 9.9b  Iceberg signature observed in altimeter elevation profiles on 10th Feb 1988.

Figure 9.9  Giant tabular iceberg A-20A, which calved from the Larsen ice shelf around March 1986, shown on (a) Navy/NOAA ice charts, (b) Revealed in altimeter height data. South Georgia is visible in both pictures.
9.4.2 Measurement of iceberg freeboard

Figure 9.10a shows, the elevation profile in the vicinity of the iceberg, at a rate of 10Hz. As the altimeter reaches the icebergs the return waveforms become distorted and data for one frame is consequently eliminated by the data editing performed on the data set and mentioned previously in section 8.1. The altimeter does however maintain lock, providing elevation measurements over the iceberg until falling back down over the open ocean.

Figure 9.10b shows an elevation profile from a collinear track from the 24th ERM cycle. By subtracting this profile from that obtained over the iceberg the residual elevation profile, shown in figure 9.10c is produced. By subtracting the two profiles, variations in the geoid can be removed. The offset present between the two profiles is caused by differences in the orbit error, tides and atmospheric correction. Over such a short distance, changes in these biases will be limited to a few cm. By subtracting the mean residual over the ocean from the profile over the iceberg and accurate measurement of the iceberg freeboard can be made.

9.4.3 Changes in the mean freeboard of giant tabular icebergs determined by the Geosat altimeter

Five further transects of both icebergs were located in a total of 48 ERM cycles examined. The resulting residual elevation profiles are shown in figure 9.11 for Larsen A and in figure 9.12 for Larsen B. The measurements of the freeboard of the two icebergs are summarised in figure 9.13, which shows the minimum, mean and maximum elevations observed.

The mean freeboard of Larsen A shows a well behaved downward trend from February 1987 to March 1988. The exception being two observations in June 1987 where a slight apparent increase in freeboard is observed. Observations of the freeboard of Larsen B from March to August 1987 show no discernable trend in the freeboard change and the overall variation is significantly greater than for Larsen A. The fact that the freeboard of Larsen B is significantly lower than that of Larsen A and that its freeboard shows a larger variation suggests that it originates from the seaward edge of the ice shelf whilst that Larsen A originates from the interior. This is commensurate with the fact that ridges, along which the fracture of the original large iceberg is likely to have occurred, were observed to run parallel to the ice shelf margin by Ridley et al. [1989].
Figure 9.10 Elevation profiles of two collinear passes with and without the iceberg present. By subtracting the two profiles, variation due to the short wavelength geoid is removed to allow the iceberg freeboard to be plotted.
Figure 9.11 Freeboard profiles of Larsen A iceberg. The horizontal offset is arbitrary and all plots are in metres.
Figure 9.12 Freeboard profiles of Larsen B iceberg. The horizontal offset is arbitrary and all plots are in metres.
The wide variation of freeboard measurements of Larsen B brings to light the problem of knowing the iceberg orientation with respect to the satellite track, particularly where a large variation in thickness occurs. This could be solved by use of co-incident imagery, a possible aim for future work. However it is not impossible that, whilst resident in the cold waters of the Weddell Gyre, the freeboard of such icebergs might actually increase due to sea water freezing onto the bottom [M. Kristensen, personal communication].

![Figure 9.13a Larsen A minimum, maximum and mean freeboard](image)

![Figure 9.13b Larsen B minimum, maximum and mean freeboard](image)

The freeboard observations for Larsen A show a much smaller deviation of surfaces heights than for Larsen B, with the minimum elevation for February 1987 significantly higher than the
maximum for March 1988. We can be fairly confident that the changes observed are not solely caused by changes in the iceberg orientation, and must partly be caused by iceberg melting. Assuming that the changes in orientation average out the observed ice freeboard changes from 33.8 to 25.5 metres over a period of 13 months. Using values give by Orheim [1980] this represents a change in thickness from 205 to 137 metres, a decrease of 68 metres. Excluding the last data point, where a substantial fall is observed, the decrease over 12 months is approximately 35 metres.

Estimates of the melt rate of icebergs vary widely but it useful to make some comparisons of this result with other observations. Huppert [1980] gives the melt rate of icebergs as:

\[
\frac{dZ}{dt} = \frac{4}{3} C \Delta T^{1.6} D^{0.25}
\]

where \( Z \) is the total iceberg thickness, \( \Delta T \) is the temperature difference between the surrounding ocean and the freezing point of sea water, \( D \) is the depth of the vertical ice wall and \( C \sim 7.5 \times 10^{-4} \text{ deg}^{-1.6} \text{ mm}^{5/4} \text{ s}^{-1} \).

The only unknown in this equation is the temperature difference \( \Delta T \) at the base of the iceberg. Figure 9.14 shows the potential temperature at depths of 100 and 200m from Gordon and Molinelli [1982] along with the positions of Larsen A on each of the six occasions where it was observed in altimeter data. The area in which the two icebergs are observed is close to the boundary of the Weddell Gyre and the Antarctic Circumpolar Current (ACC). At a depth of 100m the Antarctic intermediate water results in temperatures between -1 and 0°C, whilst at 200m the warm saline ACC water spreading towards the pole exerts an influence resulting in temperatures between 0 and 1°C.

<table>
<thead>
<tr>
<th>Water temperature (°C)</th>
<th>( \Delta T )</th>
<th>Melt rate (ma(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>-1</td>
<td>1</td>
<td>9</td>
</tr>
<tr>
<td>0</td>
<td>2</td>
<td>27</td>
</tr>
<tr>
<td>1</td>
<td>3</td>
<td>52</td>
</tr>
</tbody>
</table>

Table 9.1 Melt rate of Larsen A assuming an ice wall depth of 150m
Figure 9.14a Southern Ocean 100m potential temperature from Gordon and Molinelli [1982]. The positions of Larsen A are shown (*) although the line joining the sequence does not necessarily represent the track of the iceberg.

Figure 9.14b Southern Ocean 200m potential temperature from Gordon and Molinelli [1982]. The positions of Larsen A are shown (*) although the line joining the sequence does not necessarily represent the track of the iceberg.
Estimates for the melting rate of Larsen A are given in table 9.1 using an ice wall depth of 150m and assuming that the freezing point of seawater is -2°C. Comparing the values in table 9.1 with the observed change in ice thickness of 35m (excluding the last data point) we can see that the observations lie within the bounds suggested by theory.

This result demonstrates that monitoring the passage and deterioration of giant Antarctic tabular icebergs is feasible using data from satellite altimeters. Although the ground track sampling is not ideal at latitudes of 60° or so, near the latitude limit the probability of iceberg interception is considerably better. The ERS-1 satellite will provide coverage of the entire southern ocean and will be able to monitor icebergs soon after they calve off the Ross and Ronne-Filchner ice shelves.

9.5 Conclusions

The aim of work described in this chapter was to explore the potential of altimeter measurements of elevation over sea ice areas. The only previous study of this kind was carried out by Stanley et al. [1980] (see section 2.4.4.2) whose analysis was compromised by instrumental errors which occur when the return waveform shapes deviates from that observed over the open ocean. By retracking waveforms and eliminating bad data, it has been demonstrated that the overall RMS noise can typically be reduced from greater than 1m to less than 0.5m. This technique has been applied to Seasat data, and shows that significant geoid features can be mapped under sea ice cover.

Altimeter measurements of surface elevation in regions where low power diffuse returns exist, caused by giant multi-year ice floes or compacted ice, have shown significant increases in surface elevation, commensurate with changes in ice freeboard. When ERS-1 data becomes available, permitting coverage into the high Arctic, the monitoring the freeboard (from which thickness may be inferred) of giant multi-year floes will become a real possibility. If such a data set could be provided by satellite altimetry, it would allow for significant advances to be made in the areas of model validation and in climate monitoring (as discussed in chapter 1).

In an extreme application of the measurement of the freeboard of floating ice, the elevation of two giant tabular icebergs has been monitored over a period of greater than one year. This is
the first time that the bottom melting of giant tabular icebergs has been measured from space. Knowledge of the melting rates of such icebergs allows for better prediction of the survival time in the open ocean and, when combined with imagery, provides quantitative information on the amount of freshwater released into the ocean. The measurement of the basal melting rates of large ice masses, may also provide information of relevance to the response of the major ice shelves to changes in ocean temperature.
CHAPTER 10 CONCLUSIONS

10.0 Introduction

In this chapter we assess the contributions that the work described in this thesis have made to the study of radar altimeter data over sea ice. We then recommend direction for future work and look at the prospects of new results from the ERS-1 altimeter.

10.1 Assessment of achievements

In this section we assess the progress made in the study of radar altimeter data over polar ice through the work presented in this thesis. As a start we reproduce the objectives outlined in chapter one and assess to the extent to which they have been fulfilled.

(i) To gain a clear understanding of the operation of past and present space-borne altimeters over sea ice and the development of processing techniques to reduce the errors that occur.

- The problems of the operation of space-borne altimeters has been thoroughly addressed previously by Rapley et al [1983; 1987]. In this work processing techniques have been defined to correct for such errors and to edit data points for which precise measurements of surface elevation cannot be extracted. By employing this processing on altimeter data over sea ice the overall noise level on altimeter elevation measurements can be significantly reduced (section 9.3.1) and the RMS deviation of individual collinear tracks can be reduced by better than a factor of two.

- The parameters defined in chapter 5 show considerably more sensitivity to changing sea ice conditions (section 7.2) than parameters used by some previous researchers such as the Geosat Ice Index (section 2.4.4.9). the parameters SIGPK, SIGTD and SIGTR (defined in section 5.7.3) also show clear delineation of different zones within individual tracks (section 6.2) and also on a more global scale (section 6.4).
To advance the understanding of the mechanisms governing the response of radar altimeters to sea ice surfaces through consideration of physical theories.

In chapter 4 the theory and observations of radar backscatter from sea ice was considered. Comparison with the observations in later chapter leads to the following conclusions:

- Comparing the variation in di-electric constant given in table 4.1 with the histogram of global peak backscatter values over sea ice shown in figure 6.9 it is clear that the primary cause of variations in altimeter observations of sea ice is changing surface roughness.

- Surface measurements of the backscatter coefficient over sea ice (figure 4.9) are much lower than is typically observed over sea ice in altimeter data (chapter 6). The only occasions where backscatter levels comparable to ground measurements are observed is over vast floes and fast ice where 100% ice cover exists within the altimeter footprint.

- In section 4.7 the different components of a sea ice covered ocean were considered in terms of their surface roughness. Measurements of the small scale roughness of sea ice showed that the high power, peaked returns in altimeter data over sea ice were likely to originate from new ice (section 4.7.3) and from meltponds (section 4.7.5), but not from even smooth first year ice floes (section 4.7.4).

- Areas of smooth open water between ice floes were shown to be important both by the observations shown in figure 4.9 and by the theoretical predictions and observations of ocean surface roughness given in section 4.7.1. The theory also showed that the prevailing wind speed and dimensions of open water areas is important in determining the backscatter due to smooth water areas. In section 7.2 it was demonstrated that the backscatter coefficient is high over a compact ice edge, with a small inter-floe distance, but fell significantly where large (>km) areas of open water existed. These observations support the argument that the level of backscatter observed in the altimeter data may be dependent on the distance over which the wind can act to regenerate surface waves.

To further the interpretation of satellite radar altimeter data over sea ice through comparisons of altimeter data with sea ice climatology and by comparison with other sensors.

- The comparison of altimeter and airborne SAR data (section 7.5) showed that highly peaked waveforms seem to correspond to a regime of well broken ice.
cover adjacent to the ice edge, with a floe size limited to tens of metres, whilst the
more variable, lower power backscatter values, occur where the ice cover
becomes inhomogeneous on the scale (~kilometres) of the altimeter footprint.
This type of zonation is observed throughout the Southern ocean (section 6.4)
with highly peaked waveforms occurring around the periphery (roughly
corresponding to the zone of metre size floes given in table 2.2), and lower
power less peaked returns further into the ice pack, particularly in the areas where
ice survives throughout the year (figure 2.5a).

- The comparison of AVHRR with data from the Geosat altimeter presented in
chapter 7 showed clear delineation of features observed in both sensors.
Comparisons over compact ice boundaries showed that the radar altimeter
returns show a sharp increase in backscatter and change in pulse shape at the
boundary shown in AVHRR imagery (section 7.2.1). Over diffuse ice boundaries
the altimeter also shows a significant change from open ocean values of peak
backscatter and waveform shape although the ice boundary is much less clearly
defined. In particular where bands of ice exist the backscatter value is observed to
fall, most probably due to the regeneration of small wind waves over a large fetch.
In section 7.3 a strong correlation is observed between areas of high backscatter
observed in the altimeter data and areas of new thin ice and leads observed in
AVHRR imagery. Over vast ice floes and fast ice the altimeter exhibits 'ocean' like
returns with a lower level of backscatter than is normally observed over the ocean.

- The comparison of medium (~km) resolution infra red and visible imagery with
radar altimeter data is clearly of great value in determining the response of both
sensors to different conditions of sea ice cover. The altimeter and ATSR sensors
on board the ERS-1 satellite will provide synoptic coverage over sea ice,
overcoming the problems of geo-locating data sets from different satellites.
Further analysis of such data from ERS-1 will allow significant progress to be made
in data interpretation, and may point the way towards routine multi-sensor
products over sea ice for operational applications.

(iv) To examine how data from satellite radar altimeters can complement observations from
other sensors in contributing to applications outlined in chapter one.

In chapter 2 we assessed the capability of space-borne sensors in providing parameters
needed for climate, glaciological and operational applications. An additional aim
addressed in chapter one was the extension of satellite altimeter measurements of sea
surface height into areas covered by sea ice. In this section we review the capabilities of
satellite borne radar altimeters in the light of research presented in this thesis.

- Ice boundary location Although previous work has shown that the return from
radar altimeters changes significantly, few studies have directly compared
altimeter data with that from other sensors. In this work we have demonstrated that the altimeter responds strongly to compact ice boundaries, but in a less well defined way to diffuse boundaries (chapter 7). The work presented in chapter 8, although carried out using a data set of limited quality, demonstrated that satellite altimetry can provide valuable information on seasonal and inter-annual variations in Antarctic sea ice extent. The comparison with passive microwave observations is particularly interesting and should be a goal for future research.

- **Ice concentration** Work aimed at determining ice concentration using satellite radar altimeter data has been reported by Drinkwater [1987]. Close examination of applicable backscatter theories, particularly the two scale model (section 4.5.4), show that unambiguous measures of sea ice concentration may not be possible using radar altimetry alone. In any case the measurement of sea ice concentration is reasonably satisfied by data from passive microwave instruments.

- **New ice leads and open water mapping** The problem of mapping areas of new ice and open water within the pack is considered separately from that of measuring ice concentration, although the problem is really one of measuring the percentage of ice at high concentrations (>90%). In chapter 1 it was stated that the ocean-atmosphere heat flux over the majority of the ice pack is controlled by the small percentage of open water and thin ice (sections 1.2.1, 1.3). In chapter 2 it was stated that passive microwave observations may contain errors of one or two percent at high concentrations, limiting the use of such data in assessing the total ocean-atmosphere heat flux in the polar regions. Comparisons between altimeter data over new ice and leads presented in chapter 7 and the discussion in chapter 4 demonstrate that radar altimeters are extremely sensitive to the presence of new, thin ice and leads. Altimeter observations of the sea ice pack may therefore be able to make a contribution to global estimates of the open water percentage deep within sea ice packs. Certainly the altimeter should be able to distinguish between areas where 100% coverage of thick ice floes exist and areas where a small percentage of smooth water and thin ice exists.

- **Ice type** Analysis presented in this thesis demonstrates that satellite radar altimeter observations over sea ice covered areas are normally dominated by smooth areas of new ice or open water between floes. Only where ice floes approach the size of the altimeter footprint or where ice is compacted with very little open water may determination of ice type be possible. Certainly the possibility of ice type classification suggested by Ulander [1988], who shows backscatter over fast ice to be greater than 30dB, is at odds both with the theoretical discussion of chapter 4 and the observations of chapter 7.
Ice roughness: Measurements of ice roughness are important for determining friction values between sea ice, ocean, and atmosphere, important for sea ice dynamics. Sea ice ridging also yields information on the age and history of sea ice floes and is important in operational applications. Data presented in section 7.4 showed that direct measurement of ice floe surface roughness is possible where the altimeter footprint is 100% ice covered. Although snow cover and penetration of the radar signal may affect measurements, these are likely to be considerably less than is true over ice sheets [Ridley and Partington, 1988]. Even if only selective measurements are possible from satellite altimeters, such data may be the only that are available in some areas. In particular, measurements of sea ice roughness in the Antarctic are very sparse (section 2.2.1).

Ice freeboard: Sea ice thickness is a primary variable needed for climate monitoring (section 1.1.6), and for climate modeling. Satellite radar altimetry, at present, provides the only means for direct measurements of ice freeboard, which are closely related to ice thickness. Results presented here suggest that where diffuse returns originate from areas of 100% ice cover that high precision elevation measurements may be obtained. In chapter 6, positive height differences have been correlated with areas of old multi-year ice, which results in low power diffuse returns. In the high Arctic, vast floes are expected to occur frequently and the coverage of ERS-1 may allow measurement of sea ice freeboard variations over a wide area of the Arctic basin.

Fast ice boundary: Results from the Geosat AVHRR comparison show a consistent mapping of the fast ice boundary by the radar altimeter. In addition, different zones within the fast ice are frequently detected, suggesting detection of storm breakup and refreezing of fast ice during the winter.

Sea surface topography: Measurements of sea surface topography in sea ice areas. The analysis presented in chapter 9 showed that altimeter derived elevation data in sea ice covered areas can be significantly improved by retracking and data editing. The mean sea surface map of the Southern Ocean shown in figure 9.5b clearly delineates bathymetric features. Although in the Southern ocean, the majority of sea ice melts back during the summer, substantial areas of geophysical interest, particularly continental shelves, remain ice covered throughout the year. With the launch of ERS-1, coverage by space-borne altimetry will extent far into the Arctic basin, a geophysically important area as noted by Anderson et al [1988]. The noise on raw elevation measurements shown in figure 9.4b shows that large areas will suffer from contamination by sea ice. The correction of altimeter data over sea ice is therefore an important goal which has been advanced by the work presented in chapter 9.
To explore the geophysical applications of satellite radar altimeter data which is currently available.

Several new geophysical results have been produced from the analysis presented in this thesis:

- First map of corrected altimeter data used to produce sea surface topography underneath sea ice in the Southern ocean (section 9.3.2).
- First observation of the seasonal and inter-annual variation in Global sea ice extent using satellite radar altimetry (section 8.5).
- First measurement of iceberg melting using satellite radar altimeter data (section 9.4.3).
- First observation of sea ice (metre) scale roughness from space (section 7.4).
- First observation of variation in sea ice freeboard observed using a space-borne sensor (sections 7.4 and 9.3.3).

10.2 Directions for future work

10.2.1 Models for radar backscatter over sea ice

In chapter 4 a thorough review of radar backscatter theories applicable to sea ice was carried out. Although some attempts have been made to fit models to the altimeter response to sea ice (e.g. Brown [1982]; Ulander [1987b]) these have met with limited success. The discussion in chapter 4 and the observations made during the Lance cruise (Appendix A) show that near normal incidence backscattering from sea ice is an extremely complex phenomenon. Few measurements or models of the roughness of a sea ice covered ocean exist to determine whether current models are valid. At present the prospects for a comprehensive model of altimeter data over sea ice seem remote. Much more data on the small (mm) scale roughness of sea ice, such as that carried out by Johanssen [1988] (section 4.7.3), is needed before significant progress can be made in this area.

10.2.2 Altimeter AVHRR comparison

Although modelling of the radar altimeter response to sea ice is problematic much progress can be made from a semi-theoretical approach to the problem. For example the discussion in
Chapter 4 concluded that radar backscatter over mature ice floes was unlikely to be much higher than that observed over the ocean and would result in minimal fall off in return power within the altimeter range window. Looking at the results in chapter 7 we see that this is indeed the case over vast floes and fast ice. Another example is that of diffuse ice edges, where the idea that regeneration of wind waves in wide leads would lead to a fall in return power. An example of this was also shown in chapter 7. Further analysis should be carried out to examine the correspondence between the two sensors under different regional and seasonal conditions.

10.2.3 Altimeter mapping of sea ice extent

The results of chapter 8 clearly demonstrate the ability of satellite radar altimeter to provide measurements of the total sea ice extent, a possible indicator of global warming predicted in chapter 1. As a first step a revised study should be carried out using a data set containing both 'ocean' and 'land/ice' GDR data to overcome the problem of data editing (section 5.1.2). The next step should be a comparison of sea ice extent mapped by passive microwave and altimeter instruments over a period of at least a year. Such a study should provide valuable insight into the limitations of both sensors.

10.2.4 Altimeter measurements of surface elevation in sea ice areas

The results presented in chapter 9 showed that, by retracking and editing of data, a significant reduction in error could be achieved for data from the Seasat altimeter. Now that significant amounts of data are available from the Geosat satellite during periods of minimum ice extent a more detailed assessment of retracking over sea ice may be carried out. Techniques, such as cross-over analysis [Zwally et al., 1987c], may allow a further reduction in the RMS error of 50 cm quoted for the orbit employed in the analysis presented in chapter 9.

The measurement of the freeboard of vast ice floes should also be pursued by examining near repeat tracks from Geosat over areas of seasonal sea ice. By eliminating the effects of short wavelength geoid undulations, measurements of the freeboard of such floes, in a similar manner to that employed for the iceberg freeboard measurement (section 9.4), should be obtained. The calving of the icebergs observed in chapter 9 is a relatively rare event. Significant further observations of iceberg freeboard are therefore unlikely, although a survey of ice charts and available data should be carried out.

10.3 The promise of ERS-1

The ERS-1 satellite, due to be launched in September 1990, will provide coverage up to a latitude of ±82°. In this section we assess the possible scientific applications of data over sea ice from the radar altimeter carried on board that satellite. Data from the ERS-1 satellite will provide synoptic coverage of sea ice cover with altimeter and infra-red imaging instruments.
(ATSR). In addition the ATSR microwave segment will provide point measurements along track of the passive microwave intensities of the surface. By comparing synoptic data from these three different sensors significant progress should be possible in the interpretation of each. We now consider specific scientific projects to be carried out using data from ERS-1.

As outlined by Anderson et al. [1988] sea floor bathymetry in the Arctic basin is of considerable interest since it represents one of the youngest areas of the ocean floor in geological terms. The analysis presented in chapter 9 clearly shows that a significant fraction of the Arctic basin will remain covered in sea ice throughout the year. By employing simple retracking of altimeter data, precise measurements of sea surface elevation may be achieved where reflection from the ocean surface can be positively identified. Fortunately the design of the ERS-1 altimeter should overcome some of the problems encountered within data from the Seasat and geosat altimeter, in particular the 'height glitch' phenomenon. Analysis presented in previous chapters shows that in many cases the altimeter return originates from areas of smooth water between floes, although the presence of meltponds on the surface of thick ice floes may present some problem. Although the presence of thick sea ice cover may prevent mapping of the Arctic geoid in a single pass, the constant motion of the Arctic sea ice should allow near complete mapping over a period of time.

Observations of the changing sea ice thickness in the Arctic Ocean have recently been cited as possible evidence for global warming [New Scientist, 1989]. The coverage afforded by ERS-1 (up to ±82° latitude) should allow a much larger population of vast or giant multi-year floes and areas of compact ice than was possible with either Seasat or Geosat. If an accurate map of mean sea surface topography is produced for sea ice areas at high latitudes, as outlined above, it should be possible to measure statistics of ice floe freeboard using the same technique as applied to icebergs in section 9.4. If achieved this would be the first time that a comprehensive survey of sea ice thickness has been carried out from space.
REFERENCES


219
REFERENCES (continued)


Carsey, F.D., 1982, 'Arctic sea ice distribution at end of summer 1973-76 from satellite microwave data', JGR, 87, C8, pp. 5809-5835.

Carsey, F.D., 1985, 'Summer Arctic sea ice character from satellite microwave data', JGR, 90, C3, pp. 5015-5034.


REFERENCES (continued)


Gloerson, P. and Campbell, W.J., 1988, 'Variations in the Arctic, Antarctic, and global sea ice covers during 1978-1987 as observed with the Nimbus 7 Scanning Multichannel Microwave Radiometer', JGR, 93, C9, pp. 10666-10674.


REFERENCES (continued)


Johanssen, R., 1988, 'Laser based surface roughness measurements of snow and sea ice on the centimeter scale ', Chalmers Univ. of Tech., Gothenburg, Sweden (Research report no. 162)


Kim, Y.S.,1984, 'Theoretical and experimental study of radar backscatter from sea ice', University of Kansas, Lawrence, KS, PhD dissertation.


REFERENCES (continued)


REFERENCES (continued)


NASA, 1979, ICEX : Ice and Climate experiment, Goddard Space Flight Centre, Greenbelt, MD.


Parkinson, C.L. and Washington, W.M., 1979, 'A Large-Scale Numerical model of Sea Ice', JGR, 84, C1, pp. 311-336.


REFERENCES (continued)


Thomas, R.H., 1984, 'Observing polar regions from space', in Recent advances in Civil space remote sensing, SPIE Vol. 481, pp. 165-171.


Ulander, L.M.H., 1987a, 'Active microwave remote sensing for sea ice parameters', Chalmers Univ. of Tech., Gothenburg, Sweden (Tech. report 48L).


REFERENCES (continued)


Wadhams, P., 1983a, 'Arctic sea ice morphology and it’s measurement', J. Soc. Underwater technology, 9, no.2.

Wadhams, P., 1983b, 'A mechanism for the formation of ice bands', JGR, 88, C5, pp. 2813-2818.


APPENDIX A OBSERVATIONS OF SEA ICE DURING THE LANCE CRUISE

A.0 Introduction

The Norsk Polarinstitutt summer cruise left Hammerfest on the 17th of July 1986 aboard the 'Lance' returning to Tromso on the 13th of August 1986. The track of the Lance during the cruise is shown in figure A.1. The first week of the cruise was spent performing oceanographic measurements and included two attempts to retrieve ocean transponders. Two cruise members were also landed on 'Half moon island' (N76°59'E22°51') to do some topographic surveys.

Measurements on the ice were made at each 'ice station'. During an ice station the boat moored along side an ice floe of between 40-100 meters in size, i.e. roughly the same size as the boat for a period of 3-18 hours. The first ice station was made on the 25th July at N77°40', E19°14' in an area of fairly sparse ice cover (<1/10). After this the boat continued up the west coast of Spitsbergen picking up the helicopters and more cruise participants including Torgne Vinje at Ny Alesund. The ice edge proper was reached on the 28th July and the second ice station was taken at N81°06', E16°15'. After this the boat continued eastwards towards Kvitoya where a new weather station was erected. A total of nine ice stations were made before returning to Ny Alesund on the 9th of August.

Sea conditions were generally calm with the highest significant waveheights being encountered being of the order of 2 meters and usually much less. Air temperatures on the ice varied from -2°C to +5°C. Weather conditions were generally overcast or misty for all but two of the ice stations where conditions were clear with bright sunlight. Fog frequently occurs in coastal areas due to warm air coming off the land and encountering the relatively cold sea ice.
Figure A.1  Cruise track of the R/V Lance during the summer cruise to the Barents sea, 1986 [Rudels, 1986].
A.1 Lance cruise general observations

A.1.1 Water roughness

Photographs were taken at most opportunities where the ship was stationary and where a wind speed measurement could be made.

The presence of meltwater from nearby glaciers had a marked effect on the ocean surface roughness. This results from a strong density gradient in the top surface layer which causes selective damping of surface waves. Plate A.1 shows the ocean surface close to glaciers near the Spitsbergen coast. Although the longer wavelength components remain the small capillary waves have been damped out giving the surface a 'glassy' appearance.

Plate A.2 clearly the damping of capillary waves by a piece of ice a few meters across. Wind waves clearly visible upwind from the ice are damped downwind. As the along wind distance from the ice increases the wind waves start to reappear.

A.1.2 Sea ice cover

The main sea ice edge appeared to occur above about N80° although some patches of sea ice were seen as low as N78°. Patches several kilometres wide were quite common with fairly well defined edges and a few bits of ice seen floating in between. Plate A.3 shows an area of almost continuous but 'rotten' ice cover. The surface is covered by a large percentage of connected meltponds some connected to the ocean surface via small meltponds. Such an ice field can present problems to satellite borne instruments which estimate ice concentration. Any method relying on sensing only the first few mm of the surface would clearly underestimate the ice concentration in such a region.

Plate A.4 shows an area of open ocean (several thousand meters deep) where a low wind speed and sea ice has resulted in an almost perfectly flat water surface.
Plate A.1 Ocean surface near the Spitsbergen coast. A top layer of meltwater from the nearby glaciers apparently causes a smooth glassy surface [N76°56' E14°17' 24/7/86].

Plate A.2 Attenuation of capillary waves by ice 'bits'. The damping and slow regeneration of wind waves downwind from the ice can be clearly seen [N76°36' E19°11' 24/7/86].
Plate A.3  Area of nearly continuous 'rotten' ice cover. Large meltponds cover the surface with dark meltholes visible in some places [N80°07' E2°17' 6/8/86].

Plate A.4  An area where high ice concentration and low windspeed [W10 ~3m/s] have resulted in a very flat ocean surface. A large 2 km long floe with a freeboard ~5m can be seen in the background. [N80°07' E2°17' 6/8/86].
A.1.3 Ice floes

The ice floes seen varied in size from a few meters to up to 100 meters in diameter. Ice floes visited ranged from winter (i.e. first year) ice, with a thickness of around 1 m to multi-year ice with a thickness of nearly 5 m. The surface of the ice floes usually consisted of mm sized ice crystals consolidated into a solid surface due to apparent melting and refreezing of the surface. In general the small scale roughness of even smooth first year floes exceeded the maximum roughness for coherent reflection to occur over a significant distance, as seen on plate A.5.

A.1.4 Meltponds

Most of the ice floes visited had a number of large meltponds. Typically a 100m diameter ice floe would have 10 meltponds with a diameter larger than 5 m. Plate A.6 shows a meltpond with a much smoother surface than the surrounding water. Plate A.7 shows an ice floe where the meltponds were frozen. Plate A.8 is a close up view of one of the smaller ponds. The cm grid shows the surface to be flat to probably less than 1 mm.
Plate A.5  Small scale roughness of a smooth first-year floe. Deviation of the surface is observed to be significantly greater than the 3mm allowed by the Rayleigh criterion for the surface to be coherent.

Plate A.6  The water roughness on the meltpond, seen in the foreground, is clearly less than for the surrounding ocean.
Plate A.7  Frozen meltponds on a second year ice floe. The sled seen is approximately two meters long [N80°09' E2°05' 8/8/86].

Plate A.8  Close up of a small frozen meltpond shown in plate B.7 showing the surface to be flat within one or two mm.
A.2 Ice floe surface roughness measurements

Measurement of the large scale roughness was carried out on most ice stations using a theodolite. Observations were made over linear transects by observing the elevation of a long ruler resting on the surface. Figure A.2 shows the height profiles, height pdf and auto-correlation functions for a multi-year ice floe and figure A.3 shows the same thing for a first or second year ice floe. The first two transects were taken on a rough multi-year floe and the second two transects were taken on a smoother first or second year ice floe.

The height distribution pdf's vary between one ice floe and the next and none closely resembles a gaussian distribution. One reason for the variability of the profiles is that some profiles contained large ridges (e.g. A.2a). Meltponding can on the other hand can cause the surface height for some areas to be less than would be expected for winter ice.

The (one dimensional) rms slope was also computed along with the standard deviation in surface height. We can compute what the correlation length $l$ would be from these two parameters, if we had a gaussian height distribution and auto-correlation function, using 2.18. These three values are tabulated in table A.1.

<table>
<thead>
<tr>
<th>Floe (transect)</th>
<th>$\sigma$(m)</th>
<th>RMS slope</th>
<th>$l$(m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MY(a)</td>
<td>0.068</td>
<td>0.119</td>
<td>0.81</td>
</tr>
<tr>
<td>MY(b)</td>
<td>0.132</td>
<td>0.153</td>
<td>1.22</td>
</tr>
<tr>
<td>FY(a)</td>
<td>0.081</td>
<td>0.065</td>
<td>1.76</td>
</tr>
<tr>
<td>FY(b)</td>
<td>0.058</td>
<td>0.041</td>
<td>2.00</td>
</tr>
</tbody>
</table>

Table A.1 Ice floe surface roughness measurements.
Figure A.2 Elevation profiles over two multi-year floes.

Figure A.3 Elevation profiles over two first-year floes.
A.3 Meltpond observations

The observations described in A.1.4 show that, in some circumstances, meltponds may be important in determining the backscatter from sea ice. Measurements of meltpond radius were carried out using a tape measure and elevation measurements were carried out using a theodolite as for the large scale ice floe roughness measurements. Measurements were made on almost all the ice floes visited although usually there were insufficient numbers to extract detailed statistics concerning area and elevation for an individual ice floe. Table A.2 shows the results for one ice floe where 13 meltponds occurred.

Most meltponds observed were not circular in shape and the measurements carried out were of the approximate minimum and maximum dimensions. A mean radius was obtained by averaging the minimum and maximum dimensions.

<table>
<thead>
<tr>
<th>Meltpond</th>
<th>Dimensions (m)</th>
<th>Mean radius (m)</th>
<th>Area (m²)</th>
<th>Elevation (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>5.1 x 9.9</td>
<td>3.75</td>
<td>44.2</td>
<td>0.18</td>
</tr>
<tr>
<td>2</td>
<td>1.2 x 3.5</td>
<td>1.17</td>
<td>4.3</td>
<td>0.18</td>
</tr>
<tr>
<td>3</td>
<td>5.6 x 2.6</td>
<td>2.05</td>
<td>13.2</td>
<td>0.18</td>
</tr>
<tr>
<td>4</td>
<td>2.4 x 2.4</td>
<td>1.2</td>
<td>4.5</td>
<td>0.16</td>
</tr>
<tr>
<td>5</td>
<td>7.0 x 7.0</td>
<td>3.5</td>
<td>38.5</td>
<td>0.15</td>
</tr>
<tr>
<td>6</td>
<td>4.8 x 2.3</td>
<td>1.77</td>
<td>9.9</td>
<td>0.16</td>
</tr>
<tr>
<td>7</td>
<td>0.8 x 1.3</td>
<td>0.52</td>
<td>0.8</td>
<td>0.22</td>
</tr>
<tr>
<td>8</td>
<td>4.6 x 3.5</td>
<td>2.02</td>
<td>12.9</td>
<td>0.22</td>
</tr>
<tr>
<td>9</td>
<td>2.7 x 2.7</td>
<td>1.35</td>
<td>5.7</td>
<td>0.22</td>
</tr>
<tr>
<td>10</td>
<td>1.8 x 3.2</td>
<td>1.25</td>
<td>4.9</td>
<td>0.22</td>
</tr>
<tr>
<td>11</td>
<td>3.6 x 2.2</td>
<td>1.2</td>
<td>4.5</td>
<td>0.28</td>
</tr>
<tr>
<td>12</td>
<td>10.4 x 5.7</td>
<td>4.02</td>
<td>50.9</td>
<td>0.28</td>
</tr>
<tr>
<td>13</td>
<td>13.9 x 4.1</td>
<td>4.5</td>
<td>63.6</td>
<td>0.28</td>
</tr>
<tr>
<td>14</td>
<td>6.8 x 3.5</td>
<td>2.57</td>
<td>20.8</td>
<td>0.28</td>
</tr>
</tbody>
</table>

Table A.2 Meltpond radius and elevation.
Where meltponds were joined by linked by small drainage channels the elevation measurement was carried out on only one meltpond in each group. The method of measuring elevation results in an error of around 1 cm thus making it impossible to determine whether any two meltponds could act as a coherent pair of reflectors.

![Histogram of meltpond radius](image1)

**Figure A.5**  Histogram of meltpond radius

Figure A.5 shows a histogram of meltpond radius. There is some similarity between this distribution and the Raleigh distribution assumed by Brown in his facet model described in section 2.7.

![Histogram of meltpond elevations](image2)

**Figure A.6**  Histogram of meltpond elevations

Figure A.6 shows a histogram of the meltpond elevations. Although the measurements are not accurate enough to resolve whether the different groups of meltponds can act as coherent
reflectors it is clear that over all the facets cannot be assumed to lie in the same plane. We would therefore not expect the coherent component if any to have the amplitude described in the Brown model.

A.4 Conclusions

Photographs from the Lance cruise demonstrate that the surface roughness on meltponds is commonly less than the surrounding water. If the roughness scales of the meltponds and the open water lie either side of the critical 6mm rms roughness then meltponds may dominate the return. Where frozen meltponds exist such as that shown in Plate A.8 coherent reflection over metre size areas will certainly occur until the meltpond receives a covering of snow. The probability of thin new ice forming will also be greater where the distance between ice floes is smaller although this time the dependence will be less on wind speed and more on temperature.
<table>
<thead>
<tr>
<th>Term</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fast Ice</td>
<td>Ice attached to the shore.</td>
</tr>
<tr>
<td>Grease ice</td>
<td>Small platelets of ice herded by wind action.</td>
</tr>
<tr>
<td>Iceberg</td>
<td>A massive piece of ice calved from an ice wall or front.</td>
</tr>
<tr>
<td>Ice edge</td>
<td>The boundary between ice of any concentration and the open ocean.</td>
</tr>
<tr>
<td>Ice extent</td>
<td>The area enclosed by the ice edge.</td>
</tr>
<tr>
<td>Ice pack</td>
<td>Floating ice at concentrations greater than about 70%.</td>
</tr>
<tr>
<td>Lead</td>
<td>A linear area of open water within the ice pack sufficiently wide to allow navigation.</td>
</tr>
<tr>
<td>Nilas</td>
<td>Very smooth, thin new ice formed under calm conditions.</td>
</tr>
<tr>
<td>Tabular iceberg</td>
<td>Flat topped iceberg.</td>
</tr>
<tr>
<td>Vast ice floe</td>
<td>Ice floe with dimensions greater than 10km.</td>
</tr>
</tbody>
</table>
**LIST OF ALTIMETER DERIVED PARAMETERS**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>AGC</td>
<td>Automatic Gain Control - the gain applied to the received signal on radar altimeters.</td>
</tr>
<tr>
<td>ELEV</td>
<td>Raw elevation relative to the reference ellipsoid.</td>
</tr>
<tr>
<td>ELEVR</td>
<td>Retracked elevation relative to the reference ellipsoid.</td>
</tr>
<tr>
<td>GELEV</td>
<td>Raw elevation relative to the geoid.</td>
</tr>
<tr>
<td>GELEVR</td>
<td>Retracked elevation relative to the geoid.</td>
</tr>
<tr>
<td>LEDIF</td>
<td>A measure of the skewness of the leading edge of radar altimeter return waveforms.</td>
</tr>
<tr>
<td>LEWID</td>
<td>A measure of the width of the leading edge of radar altimeter return waveforms.</td>
</tr>
<tr>
<td>SIGPK</td>
<td>A measure of the peak backscatter coefficient in a radar altimeter return.</td>
</tr>
<tr>
<td>SIGTD</td>
<td>A measure of the rate of fall off in return waveforms with increasing incidence angle.</td>
</tr>
<tr>
<td>SIGTR</td>
<td>A measure of the deviation of return waveforms from an exponential fall off.</td>
</tr>
<tr>
<td>SWH</td>
<td>Significant Wave Height (= 4 x standard deviation of surface height for a gaussian height distribution)</td>
</tr>
<tr>
<td>VATT</td>
<td>Voltage proportional to ATTitude - The ratio between off-nadir and nadir returned power used by the Geosat altimeter to estimate off-pointing.</td>
</tr>
</tbody>
</table>
**LIST OF SYMBOLS**

- **a**: AGC control loop time constant
- **A**: surface area
- **A_n**: area of nth facet
- **ΔAGC**: Error on estimated AGC value used to update the altimeter gain control loop.
- **c**: velocity of light
- **E**: electrical field at a point resulting from reflection of a facet
- **D**: depth of a vertical ice wall
- **f_n**: scattering pattern of nth facet
- **F**: fraction of the surface resulting in quasi-coherent reflection
- **g**: acceleration due to gravity
- **G**: antenna gain
- **h**: satellite height
- **h'**: height rate
- **h_0**: satellite reference height
- **Δh**: height difference
- **H**: satellite height above the reference ellipsoid
- **I_n**: Modified Bessel function (order n)
- **J_n**: Bessel function of the first kind (order n)
- **k**: radar wavenumber
- **l**: correlation length
- **L_{atm}**: loss due to atmospheric attenuation
- **n**: waveform sample number
- **M**: number of areas
- **N**: number of facets
- **p_i**: power in a waveform sample corresponding to a single return pulse
- **P_{fs}**: Flat surface impulse response
- **P_i**: power in waveform sample i
- **P_t**: transmitted power
- **P_r**: received power
- **q_s**: probability density function of surface heights in terms of delay time
LIST OF SYMBOLS (continued)

\[ |IR(0)|^2 \] Fresnel reflection coefficient at normal incidence
\[ r \] radius of an aperture
\[ r_e \] radius of the earth
\[ r_{PLF} \] pulse limited footprint radius
\[ r_{RWF} \] range window footprint radius
\[ r_{BLF} \] beam limited footprint radius
\[ R \] range
\[ \Delta R \] range difference
\[ RCS \] radar cross-section
\[ s \] root mean square slope
\[ S_r(t) \] transmitted pulse profile
\[ t \] time delay
\[ \Delta t \] time difference
\[ T \] temperature
\[ \Delta T \] temperature difference
\[ T_b \] brightness temperature
\[ U_a \] wind speed
\[ U_* \] wind friction velocity
\[ V_b \] brine volume
\[ W_{3db} \] 3dB width of return waveform in samplers
\[ x \] distance
\[ X \] fetch
\[ Z \] iceberg thickness
\[ \alpha \] tracker height time constant
\[ \alpha_x \] ocean wave attenuation coefficient
\[ \beta \] tracker height rate time constant
\[ \delta \] depression angle
\[ \varepsilon \] surface emmisivity
\[ \varepsilon_A \] electric field source strength per unit area of an aperture
\[ \varepsilon_i \] relative permittivity of pure ice
\[ \varepsilon_r \] relative permittivity (magnitude of the dielectric constant)
\[ \varepsilon_{si} \] relative permittivity of sea ice
\[ \phi \] phase
\[ \Delta \phi \] phase difference
\[ \Gamma \] transmission coefficient
\[ \eta \] Average number of facets per unit area
\[ \lambda \] wavelength
\[ \pi \] Pi
\[ \omega \] frequency
\[ \theta \] angle of incidence
\[ \theta_b \] antenna 3dB width
\[ \sigma^0 \] backscattering coefficient
\[ \sigma_{QS} \] quasi-coherent component of backscatter
\[ \sigma_s \] surface component of backscatter
\[ \sigma_v \] volume component of backscatter
\[ \sigma_h \] standard deviation of surface height
\[ \sigma_p^2 \] second moment of radius distribution
\[ \tau \] transmitted pulse length
\[ \tau' \] time delay over which power is received from a single range ring over a rough surface.
\[ \psi \] azimuth
\[ \Psi \] waveheight spectrum
\[ \Psi_s \] long wavelength waveheight spectrum
\[ \Psi_l \] short wavelength waveheight spectrum
\[ \zeta \] off-pointing angle