RADAR ALTIMETRIC STUDIES OF POLAR ICE

by

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Frontispiece

This cyclonic eddy was photographed during the 1984 Marginal Ice Zone Experiment, by Dr. R.A. Schuchman of the Environmental Research Institute of Michigan. The centre of the eddy was located at approximately 78° 40’N and 2°W on 30th June. The diameter of the eddy is 30-40 km. This oblique aerial photograph was taken at an altitude of 7 km from the cockpit of the Canadian Centre for Remote Sensing (CCRS) Convair-580 aircraft.
PREFACE

The work presented in this thesis was carried out at the Scott Polar Research Institute (SPRI) between October 1984 and November 1987 under the guidance of my supervisor Dr V.A. Squire, to whom I am indebted. I am also grateful to Dr G. Rees who was my acting supervisor during the latter stages of thesis production. I thank Dr D.J. Drewry, ex-director of SPRI for his encouragement and help, and for the use of the facilities available at the Institute. The Natural Environment Research Council funded me during the three years research and Emmanuel College generously sponsored my attendance at various Conferences and Symposia and fieldwork in the Labrador Sea in March and April 1987.

The most important part of this work, the data collection, was performed in the Arctic during the Marginal Ice Zone Experiment of June and July 1984. A truly collective operation, involving scientists from all over the world took place. Gathering of altimetric data was spearheaded by a British team whose efforts in building an airborne radar altimeter were rewarded by several flights mounted in the NASA CV-990 aircraft during the campaign. I therefore am indebted to the Rutherford Appleton Laboratory (RAL) who designed and built the instrument used, and in particular Andrew Birks for making my participation in this possible. His supply of digital data tapes and helpful comments have enabled a successful and varied study to be undertaken.

For their helpful discussions I am grateful to Neil McIntyre and everyone involved in the ‘Analysis of Altimetry Data from the Marginal Ice Zone Experiment’, a study undertaken under contract for the European Space Agency in 1985. Of my colleagues at SPRI I thank Julian Dowdeswell for his inspiration and help with image processing and analysis, and Paul Cooper for help with tape reading and computer algorithm development. Both Mark Elkington and Chris Rapley of the National Remote Sensing Centre deserve thanks for time spent using the Algorithm Development Facilities, and in particular the GEMS digital image processing equipment at the Royal Aircraft Establishment, Farnborough. The Kiruna satellite receiving station and ESA Earthnet supplied the Landsat imagery.

Finally, I wish to thank Lesley Wilson for her help with drafting some of the figures and for her encouragement, and Rob Southern for his help with photographicis.

Although altimetric data collection in all cases involved other people, the reduction, analysis, and interpretation of the results, except where otherwise stated, are the original work of the author. I declare that this thesis does not exceed the regulations on length and that it has not been submitted, in whole or in part, for a degree at this or any other university.

This dissertation is the result of my own work and includes nothing which is the outcome of work done in collaboration.
SUMMARY

Active microwave sensors are known to provide valuable information regarding snow and ice surfaces in the polar regions, where darkness and cloud cover prevail. Here, data collected in the Arctic by a Ku-band microwave radar altimeter, designed and constructed in the UK, are analysed. The two main components of this study comprise data gathered in the East Greenland Sea marginal ice zone and over two Svalbard ice caps.

A systematic treatment is made of the electromagnetic properties of snow and ice at 13.81 GHz, and the differences between various polar surface media are highlighted. Theoretical and empirical models are presented which enable calculation of the relevant dielectric and scattering properties of snow and ice layers. Parametric studies are undertaken to give insight into the range of scattering conditions likely to be encountered by a radar altimeter in the regions investigated.

Examples of altimetric data and results of their analysis are presented, demonstrating the effects of different ice types and terrain upon incident altimeter pulses. Waveforms are characterised by their shape, and certain forms are linked with particular physical properties of the surface. To this a variety of supporting information is added in order to verify and validate interpretations of these results. Algorithms are proposed which enable geophysical information to be derived from altimetric data.
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ABBREVIATIONS AND ACRONYMS

Bic  Biomation interface counter
CCRS  Canadian Centre for Remote Sensing
CCT  Computer Compatible Tape
CV-990  Convair 990 aircraft
ERIM  Environmental Research Institute of Michigan
ERS-1  Earth Remote Sensing satellite no. 1
ESMR  Electronically Scanning Microwave Radiometer
HP 1000  Hewlett Packard 1000 computer
MIZ  Marginal Ice Zone
MIZEX  Marginal Ice Zone Experiment
RAL  Rutherford Appleton Laboratory
RES  Radio Echo Sounding
SAR  Synthetic Aperture Radar
SMMR  Scanning Multichannel Microwave Radiometer
SPRI  Scott Polar Research Institute
NASA  National Aeronautics and Space Administration
NOAA  National Oceanographic and Atmospheric Administration
VDU  Video Display Unit
LIST OF SYMBOLS

\( A \)  
\( A_e \)  
\( A_{ill} \)  
\( c \)  
\( d \)  
\( D_{BLF} \)  
\( D_{PLF} \)  
\( D_f \)  
\( f \)  
\( f_0 \)  
\( g \)  
\( G \)  
\( h \)  
\( H \)  
\( k \)  
\( k_e \)  
\( k_o \)  
\( l \)  
\( L \)  
\( L_e \)  
\( n \)  
\( N \)  
\( N \)  
\( P \)  
\( P_r \)  
\( P_t \)  
\( p(\cdot) \)  
\( r \)  
\( r_n \)  
\( R \)  
\( R_{12} \)  
\( R_d \)  
\( R_g \)  

snow grain diameter  
antenna effective area  
illuminated area  
speed of light \((3 \times 10^8 \text{ m s}^{-1})\)  
depth beneath surface  
diameter of beam limited footprint  
diameter of pulse limited footprint  
diameter of simultaneously illuminated footprint  
frequency  
relaxation frequency  
acceleration of gravity  
antenna gain  
height above mean surface  
aircraft height  
wave number, \(2\pi/\lambda\)  
absorption coefficient  
wave number in free space, \(2\pi/\lambda_o\)  
surface correlation length  
loss factor  
waveform leading edge slope  
refractive index, \((\sqrt{\epsilon^*})\)  
number of pulses sampled or averaged  
number of particles per unit volume  
power  
received power  
transmitted power  
probability density function  
radius of a particle  
radius of a range ring  
range to target  
power reflexion coefficient between mediums 1 and 2  
power reflexion coefficient of a diffuse surface  
power reflexion coefficient of a glistening surface
\( R_n \)  power reflexion coefficient at normal incidence
\( S_r \)  absorption cross-section of receiving antenna
\( t \)  time delay, time interval
\( \tan \delta \)  dielectric loss factor, \((\varepsilon''/\varepsilon')\)
\( T \)  temperature
\( u \)  Weiner’s ‘formzahl’ or form number
\( v \)  aircraft velocity
\( V \)  volume fraction
\( W_v \)  volume fraction of water
\( X \)  waveform trailing edge attenuation coefficient
\( \alpha, \beta \)  tracker constants (see section 2.3.3)
\( \beta \)  root mean square surface roughness
\( \gamma \)  surface tension
\( \delta_p \)  depth of penetration
\( \delta \)  antenna gain pattern correction
\( \varepsilon \)  permittivity
\( \varepsilon', \varepsilon'' \)  real, imaginary part of \( \varepsilon \)
\( \varepsilon^* \)  complex dielectric constant
\( \varepsilon_0 \)  permittivity of free space, \(8.85 \times 10^{-12} \text{ farad m}^{-1}\)
\( \eta \)  intrinsic impedance
\( \theta \)  angle of incidence
\( \theta_i \)  local angle of incidence
\( \theta_{3d} \)  full 3dB beam width
\( \lambda \)  wavelength
\( \lambda_0 \)  free space wavelength
\( \mu \)  magnetic permeability
\( \mu_0 \)  magnetic permeability of free space, \(4\pi \times 10^{-7} \text{ henry m}^{-1}\)
\( \rho \)  density of a material
\( \rho_{12} \)  amplitude reflexion coefficient between mediums 1 and 2
\( \rho_d \)  surface roughness factor, diffuse surface
\( \rho_s \)  surface roughness factor, specular surface
\( \sigma \)  conductivity
\( \sigma_b \)  backscattering radar scattering cross-section of a particle
\( \sigma_v \)  volume backscattering coefficient (radar reflectivity)
\( \sigma_h \)  standard deviation of surface height
\( \sigma_h^2 \) variance of surface height
\( \sigma^0 \) backscattering cross-section (per unit area), also called backscatter coefficient
\( \tau_{12} \) amplitude transmission coefficient between mediums 1 and 2
\( \tau \) duration of a pulse (pulse length)
\( \Upsilon \) power transmissivity (ie. \( 1 - R_{12} \))
\( \phi \) surface slope
\( \phi_x \) half beam width in x plane
\( \Phi \) phase shift
\( \chi \) dimensionless Mie size parameter, \( \chi = \frac{2\pi r}{\lambda} \)
\( \omega \) angular (radian) frequency
Ablation
All processes by which snow, ice, or water in any form are lost from glaciers, floating ice, or snow covers. These include melting, evaporation, calving and wind erosion.

Accumulation
All processes by which snow, ice, and water are added to glaciers, floating ice or a snow cover (ie. opposite of Ablation).

Brash ice
Accumulations of floating ice made up of fragments < 2 m across, the result of pulverisation of larger floes.

Capillary wave
A wave whose velocity of propagation is controlled primarily by the surface tension of the liquid in which the wave is travelling. Water waves of less than about 2 cm are regarded as capillaries.

Columnar ice
Form of sea ice in which crystals are aligned in a columnar manner, as opposed to frazil ice which displays random crystal orientation.

Congelation ice
Ice which typically forms beneath a layer of frazil ice in non-turbulent water. Crystals form and congeal in a structured manner enabling crystal growth to proceed in a predominantly vertical direction.

Crevasse
A fissure formed in glacier ice.

Crust
A hard snow surface lying upon a softer layer. Crusts may be formed by sun, rain, or wind.

Eddy
A circular movement of water usually formed between two currents flowing counter to each other, or along the edge of a permanent current. In areas of marginal ice the eddy may entrain ice floes, thus tracing the spiral motion.

First year ice
Floating ice of not more than 1 year's growth. Normal thicknesses range from 30 cm to 2 m. Characteristically smooth and level surfaces where undisturbed by pressure, but where ridges occur they are rough and sharply angular.

Frazil ice
Fine spicules or plates of ice in suspension in water, or ice which forms in supercooled water that is too turbulent to permit coagulation into congelation ice. Crystals as a result are randomly aligned.
Gravity wave
A wave whose velocity of propagation is controlled primarily by gravity. Water waves of length greater than about 3 cm are considered gravity waves.

Grease Ice
Slurry of ice crystals in suspension in the sea that gives the sea surface a greasy or glassy appearance.

Glacier
A mass of snow and ice continuously moving from higher to lower ground. The principal forms of glaciers referred to here are: ice sheets, and ice caps.

Glacierized
Ice covered land, or land presently overlaid by glacier ice; as opposed to Glaciated which means land covered in the past by glacier ice.

Gyre
The circular rotation of water in the sea, driven by the prevailing winds and the Coriolis effect.

Ice edge
The boundary at any given time between open water and sea ice. An ice edge may be termed compacted when it is clearly defined, or diffuse/dispersed when it forms an indefinite edge.

Ice floe
A piece of floating ice which formed on the surface of the sea. Floes are subdivided by size as follows: small floes 20-200 m; medium floes 100-500 m; large floes 0.5-2 km; vast floes 2-10 km; giant floes > 10 km.

Ice sheet or ice cap
A mass of ice and snow of considerable thickness and large area. Ice sheets are generally resting on rock (floating margins of ice sheets are termed ice shelves). Ice sheets of less than about 50,000 km² are called ice caps.

Ice stream
Part of an ice sheet in which the ice flows more rapidly than the surrounding ice. The margins are sometimes marked by a clear break in surface slope or crevassing.

Lead
An area of open water in a floating ice cover (vis. a navigable passage).

Marginal ice
Zone of sea ice in the vicinity of the ice edge

Multiyear ice
Sea ice which has survived the first two summers of its existence. It normally attains thicknesses of 2 m and more. It stands higher out of the water than first year ice. Summer melting rounds and smooths pressure relief to hummocks, while differential melting accentuates the minor relief.
**New ice**  
A general term for recently formed floating ice including frazil ice, grease ice, and nilas.

**Nilas**  
A thin elastic crust of floating ice which flexes upon waves and swell. It may be up to 10 cm in thickness.

**Nunatak**  
A rocky crag or small mountain projecting from and surrounded by an ice sheet.

**Outlet glacier**  
A glacier which drains an ice sheet or ice cap and flows in a valley or well defined channel.

**Pressure ridged ice**  
A term for floating ice which has undergone compressive stress and shearing. In places ice is forced upwards into ridges of ice rubble called pressure ridges.

**Rafted ice**  
A form of pressure ice in which one floe has overridden another.

**Rafting**  
Pressure process by which one floe overrides another

**Ridge**  
A ridge or wall of broken floating ice forced upwards by pressure.

**Rotten ice**  
Floating ice which has become honeycombed or roughened in the course of melting, and which is in a state of advanced disintegration.

**Sastrugi**  
Sharp, irregular ridges formed on a snow surface by wind erosion and deposition, Ridges are formed parallel to the direction of the prevailing wind.

**Sea ice**  
Any form of ice found at sea which originated from the freezing of sea water.

**Snow depth**  
The vertical distance between the surface of a snow layer and the ground or ice beneath it.

**Snow line**  
The line or zone on land that separates areas in which fallen snow disappears in summer from areas in which snow remains throughout the year. The altitude of the snow line is controlled by the temperature and the amount of snowfall.
CHAPTER 1

INTRODUCTION

1.1 The cryosphere: a case for remote sensing

The global system is seen to contain five important geophysical elements: the hydrosphere (principally the oceans), atmosphere, biosphere, lithosphere, and cryosphere. In recent years it has been repeatedly demonstrated that all are inseparably intertwined (Atlas et al., 1986). To gain a comprehensive understanding of the physical characteristics and mechanisms of interaction of these subsystems presents a considerable challenge, not only because of the complexity of the processes, but because of the difficulties in obtaining global observations at the necessary fine spatial and temporal scales. The latter problems are particularly severe in the case of the cryosphere, which until recently remained largely uninvestigated. It has become increasingly apparent that the cryosphere plays a fundamental interactive role in influencing the character and behaviour of the planet on which we live (Rapley et al., 1983).

Most of the world's ice and snow lies in the polar regions, an area of some 50 million square kilometres. The Arctic and Antarctic Ice sheets themselves constitute some 10% of the surface land area of the earth. Sea ice on average covers $24 \times 10^6 \text{ km}^2$ of the ocean surface ($\sim 7\%$), and exhibits marked seasonal and annual variability (Squire, 1984), increasing to 13% of the ocean surface at times. The size and influence of such large ice masses is so important that they have a control upon both regional and global climate, as well as eustatic changes, ocean circulation, and atmosphere-ocean interactions. Increased awareness of the role of polar ice in modulating and responding to world climate, in controlling sea level, and in modifying ocean properties has exposed a lack of understanding of ice behaviour (Thomas, 1986). It is now known that polar ice is extremely dynamic, having a continually changing character. Wherever detailed measurements have been made the ice has been seen to respond to 'forcing' in different ways, by local thickening, or changes in surface terrain characteristics. Nevertheless, such measurements are sparse due to the expense and logistical inconvenience of in situ data collection. Indeed, much of Greenland and the Antarctic remain unsurveyed, and the annual extent of the sea ice cover is not known with any great degree of accuracy.

Weather over the poles is poor for most of the year. Clouds predominate over vast areas, and for several months each winter there is no sunlight. All these factors make aerial or satellite data collection in the visible part of the spectrum as difficult as ground data collection. Such problems have been solved by using data from microwave remote sensing instruments such as the radar altimeter, synthetic aperture radar (SAR), scatterometer, and radiometer. These
instruments, whether satellite-borne or airborne, can retrieve data during periods of darkness and through most cloud and fog conditions, and are therefore well suited for this type of work. Their data provide a reliable source of multitemporal (synoptic) observations over the polar regions.

The major goal of this study is to investigate how microwave radar altimeters in particular function in the polar regions, and to improve our ability to convert altimeter measurements into useful geophysical parameters. The possibility of extracting critical glaciological information from microwave radar altimetry over both continental ice and sea ice is pertinent to the polar scientific community at large.

1.2 Background

In this section the basic principles of altimetry are dealt with as a grounding for observations made in later chapters. First a brief review is made of the historical aspects of microwave satellite radar altimetry, and of the developments in the recent literature. Many key papers have been collectively published which are outside the scope of this study, but which give extensions of the field of altimetry in other disciplines. These may be found in Gower (1981), Bernstein (1982), Kirwan et al., (1983), Allan (1983), and Robinson (1985).

1.2.1 History of satellite radar altimetry

The radar altimeter is a precise ranging instrument which has been used successfully on several satellites starting with the SKYLAB S-193 instrument in 1973 (McGoogan et al., 1974). NASA's Earth and Ocean Physics Applications Programme subsequently adopted altimetry as one of its principal techniques, with the aim of achieving 10 cm sea surface height measurements by 1980 (Dooley et al., 1974). The ensuing satellites, GEOS-3 (launched in 1975) and SEASAT (in 1978) became increasingly more sophisticated and the latter achieved the goal for ranging precision. With this precision one can measure wave heights and the properties of the geoid and detect small differences in surface height associated with the major currents in the open ocean (Gower, 1979; and Leitao et al., 1978).

From the early developments to the present, each advance has been accompanied by a burgeoning interest in altimetry, illustrated by a growing source of literature. Of note is the extraordinary diversity of suggested uses of altimetric data in Oceanography, Geodesy, Geophysics, topographic surveying, monitoring lake levels, and Glaciology. The instrument has proved to be quite remarkable from this point of view when compared with other remote sensing instruments.

Satellites with a near-polar orbit and aircraft equipped with microwave instruments have
revolutionised the study of the cryosphere. Most importantly they enable observations in areas previously poorly investigated, and in all weathers, because of their capacity to look through cloud cover. Radar altimetry has been demonstrated to provide a powerful means of studying polar ice, and processing of both GEOS-3 and SEASAT instrument data has produced impressive results (Zwally et al., 1983) to a 72° latitudinal limit over both polar regions. For example, the Seasat altimeter produced over 600,000 useable measurements over the Arctic and Antarctic terrestrial ice masses (i.e. Greenland and Antarctica), and even more useable measurements over polar sea ice during its short life (June-October 1978). When continuously tracking the altimeter yielded range measurements at approximately 0.1 second intervals, representing 662 m spacing along the subsatellite track (Brenner and Martin, 1982).

Unfortunately, to date all information collected by altimeters which is pertinent to glaciologists and polar scientists alike has been a by-product of missions flown with the objective of gathering oceanographic data. The planned launch of the European Space Agency’s ERS-1 remote sensing satellite in the early 1990’s will provide the scientific community with the first satellite mission to include monitoring of ice surfaces as an objective of the mission. Included in its sensor package will be a $K_u$ band radar altimeter equivalent to the instrument carried by Seasat in 1978. In general, such instruments of the near-future will be based on the proven technology of the highly successful SEASAT satellite altimeter. Strict design specifications and technological advances will ensure that they are similarly successful and realise their potential as geophysical remote sensing instruments. Along with Geosat and NROSS, ERS-1 will be the beginning of a new breed of civil and military satellites carrying altimeters, and it signifies the start of a new era in polar remote sensing.

1.2.2 The radar altimeter: principles of operation

This section briefly outlines the principles of operation of a radar altimeter which are relevant to this study. It discusses the function, operating modes, and the recorded echo or pulse waveform characteristics pertinent to the following chapters.

1.2.2.1 The radar altimeter concept

The familiar term RADAR is an acronym for RAdio Detection And Ranging. Originally the term was applied to instruments developed for the purpose of measuring the range to distant objects at radio frequencies. Today, however, the term is applied more loosely to describe a broad class of devices which operate in or near the microwave region of the frequency spectrum. There has been continual development of new radar capabilities and continual improvements to the technology and practice of radar. Synthetic Aperture Radar, for instance,
has sophisticated imaging capabilities, far removed from the original concept of a ranging and detection device. However, notwithstanding essential differences such as the frequency of operation, the microwave radar altimeter was also conceived as a precise ranging instrument.

Altimeters are radars which transmit short pulses of energy vertically downwards towards the earth beneath them. The return time of the pulse after reflection from the surface is measured yielding the terrain clearance of the sensor; and the envelope of power in the return pulse is recorded, giving a brief history of backscattered power from the surface. In this respect it performs the same role as the original radar, albeit with improved accuracy and increased precision.

1.2.2.2 Beam and pulse limitations

Short-pulse altimeters can operate in either of two modes, (a) beam limited, or (b) pulse limited. In the former the pulse may be viewed as being contained within a very narrow cone (representing the antenna beam width) as it progresses downwards: in this case the echo shape versus time corresponds with the probability density function of the surface height, and the mean surface height may be derived roughly from the peak of the waveform. When mounted on a satellite at an altitude of 800 km this mode may only be achieved with beamwidths of the order of 10 arcminutes or less (Rapley et al., 1983). This would require an antenna of the order of 10 m or more in diameter, and is impractical in terms of size, cost, and the problems of pointing accuracy. Thus the beam limited mode of operation has previously been ruled out.

In the latter, more common pulse limited mode, the engineering constraints are less stringent and the antenna size need only be of the order of 1 m diameter to give beam widths of the order of several degrees at a wavelength of 0.022 m (ie. 13.8 GHz). This makes the antenna unit more portable on a space platform or aircraft. Providing the transmitted pulse width \( c \tau \) is narrow, the smaller components of the surface roughness can be resolved. For \( \text{rms} \) roughnesses of 1 m to be resolved this requires the pulse length \( \tau \) to be of the order of 3 ns. In this mode echo returns are stretched in time delay by metre-scale surface perturbations, rising in amplitude as backscatter from the nadir point is received first from the crests of the surface undulations, then mid-points, and finally the troughs between them. The leading edge of the impulse response then corresponds to the integral of the specular point density as a function of range since each stratum of point scatterers adds backscattered energy at an appropriate value of delay time (Rapley et al., 1983). For surfaces with height probability distributions which approximate to a Gaussian, such as the ocean, the range to the mean surface may be derived from the time delay to the mid-point of the leading edge of the return pulse. Since historically the primary interest in altimetry has concentrated on measurements of more or less flat uniform ocean surfaces, the pulse limited mode of operation has been adopted as the norm.
The next question which arises is what area upon the surface does the altimeter sense, and what information is contained in individual data records other than accurate range measurements.

1.2.3 The return pulse limited waveform

Over a flat homogeneously rough horizontal surface, viewed at normal incidence, the return pulse shape (or impulse response) \( P_r(t) \) may be expressed as a convolution of three functions:

\[
P_r(t) = S_r(t) * P_{FS}(t) * q(t)
\]

The situation for the nadir pointing case is illustrated in Figure 1.1, where; \( S_r(t) \) is the shape of the transmitted pulse or 'point target response'; \( P_{FS}(t) \) is the radar cross-section as a function of delay time (over the horizontal surface) weighted by the antenna gain pattern \( G^2(\theta) \), or simply the 'flat surface impulse response'; and \( q(t) \) characterises the surface roughness and is the mean density of point scatterers as a function of delay time.

![Figure 1.1](image)

**Figure 1.1** Various distributions controlling the return pulse shape in the pulse limited mode of altimeter operation.
Figure 1.2 shows an altimeter transmitting short Gaussian pulses with pulse length ($\tau$) towards a flat isotropically scattering surface. A typical return pulse is shown at the base of the diagram, in which the power is proportional to the area of surface illuminated (antenna pattern effects are included later). The nominal 3 dB half beam limits of the altimeter, and the angle subtended by the periphery of the illuminated area at any instant, are indicated in the upper section, and the area illuminated by the footprint is shown in the centre section. As the spherical pulse shell propagates the returned power remains zero until a time $t_0$. This power is called the ‘first return’ and results from reception of backscatter at normal incidence from the surface the instant at which the wavefront intercepts the surface. The distance at this point of illumination is simply the sensor altitude $H$, and so the time $t_0$ is equal to $2H/c$, where $c$ is the velocity of light. Power grows to a peak at a time $t_1$, when the spot on the surface has expanded to the point at which the maximum amount of energy can be simultaneously received by the instrument. The illuminated area $A_1$, or footprint, at time $t_1$ is

$$A_1 = \pi r^2$$

where;

$$r_1^2 = d_1^2 - H^2,$$

and

$$d_1 = H + \frac{c \tau}{2}.$$  \hspace{1cm} (1.3)

Thus

$$A_1 = \pi \left( H c \tau + \frac{(c \tau)^2}{4} \right).$$  \hspace{1cm} (1.4)

For an aircraft altimeter at an altitude of around 10 km, or a satellite altimeter the second term in equation 1.5 is negligible and so $A_1$ is equal to $\pi H c \tau$, to a good approximation. Eventually, an annulus forms, and the received power is proportional to the annulus area. However its geometry is such that its area $A$ remains constant as the diameter increases. If antenna pattern attenuation is not accounted for the power $P_t$ would remain constant up until a time $t_n$ when the footprint size reaches the limits of the beam: at this point the power falls abruptly to zero. But the antenna pattern attenuation introduces an exponential decay in pulse waveform power. As the annulus expands toward the beam limits (ie. $r_1-r_6$), energy scattered back in the direction of the sensor arrives from increasing angles off-nadir (ie. angular displacement from the antenna boresight axis) and further from the region of maximum gain of the antenna (see Appendix B). In typical ocean waveforms this slowly decaying portion of return waveforms is known as the ‘plateau region’ (Barrick and Swift, 1980), because as the spherical region continues to intercept a constant annular mean surface area (as it spreads outwards from the nadir point) $P(t)$ remains at a plateau. However, most accounts of the action of pulse
Figure 1.2 Interaction between altimeter pulses of width $c\tau$ and a diffusely scattering flat surface, showing the evolution of the beam geometry and footprint annulus, along with the resulting waveform.
limited altimeters have concentrated on their performance over flat Lambertian scatterers. In later chapters this discussion is extended to cover more complex surfaces. As a consequence, although the diagrammatic explanations of the return pulse shape in Figures 1.1 and 1.2 are correct, the concept of a slow decline in power after the peak in the return, or 'plateau region' in return pulses is not strictly correct. Surfaces which approach specular reflectors or strongly directional antennas cause a more rapid decline in power. The term 'trailing edge' is thus preferred for the portion of return waveforms arriving after peak amplitude.

1.3 The Marginal Ice Zone Experiment

1.3.1 Experiment overview

The Marginal Ice Zone Experiment, or MIZEX as it is commonly known, is a programme to observe mesoscale (ie. on a scale of 10-100 km) processes at work in the region where polar air, water and ice masses interact (Wadham et al., 1981; Johannessen et al., 1983). It was established as a project which would last for several years, thus permitting field experiments and observations of one of the most important geophysical boundary zones in the Northern Hemisphere. It seeks to understand the mesoscale processes that dictate the advance and retreat of the ice margin (Johannessen, 1987). Successful modelling and parameterisation of processes which govern the location and state of the marginal ice zone (MIZ) are important not only for predicting the impact of sea ice on global climate but also furthering man's activities in the MIZ regions (Muench, 1983). There are several areas of special interest which include offshore oil exploration, seaborne transport of Arctic resources, development of fisheries close to the ice margin, and naval operations.

Various mesoscale experiments have been carried out in marginal ice zones before the conception of MIZEX, which included; BESEX (Kondratiev et al., 1975), MIZPAC (Paquette and Bourke, 1979), NORSEX (NORSEX, 1983), MIZLANT (Newton and Piper, 1981), BERING '79 (Bauer and Martin, 1980), and BERING '81, East Greenland wave-ice experiments (Wadhams, 1979), drifting buoy programmes (Vinje, 1978), and YMER-80 (Ymer, 1981). These focussed attention on the need for coordinated, multidisciplinary studies of conditions and processes near ice-open ocean boundaries. This prompted a workshop held in Voss, Norway in October 1980 to define the nature of the programmes required in a complete MIZEX.

MIZEX-West was the first of three major field efforts, and took place in February 1983 in the Bering Sea. Both shipborne and airborne observations were utilised in the study of the marginal ice zone at the time of the ice's maximum extent. The US Coastguard icebreaker Westwind and the US National Oceanographic and Atmospheric Administration (NOAA) ship
Discoverer collected surface data, and two aircraft, the NOAA P-3 and the NASA Convair 990 (CV 990) shown in Plate 1.1, provided remote sensing coverage. It was during flights made by the CV 990 that a prototype version of a microwave radar altimeter instrument designed and built by the Rutherford Appleton Laboratory (RAL) was tested. Considerable work has been carried out since MIZEX-West was undertaken and papers describing some of these analyses appeared in a special issue of the Journal of Geophysical Research (Vol. 88, No. C5, 1983).

MIZEX-East was the second major component of MIZEX, and began with a pilot study known as MIZEX '83 in June and July 1983. The Greenland Sea MIZ, north of 78° N was chosen for this experiment because it is the region where most of the heat and mass exchange takes place between the Arctic Ocean and the North Atlantic; it is therefore crucial thermodynamically. Furthermore, the ice edge combines many of the most interesting features of all marginal ice zones. The chartered sealer Polarbjørn spent 6 weeks drifting with the ice and conducting transects across the ice edge, with assistance during part of the period from the German research ship FS Polarstern and the Lance, a research vessel operated for the Norsk Polarinstitutt. Airborne remote sensing was conducted with three aircraft equipped with a variety of sensors. These demonstrated the utility of remote sensing techniques in the study of both physical and geophysical processes of the MIZ.

MIZEX '84 was undertaken the following year at the same time and place. It was the largest ever coordinated experiment conducted in the Arctic MIZ, and was established as a project which would last several years. This, the culmination of four years planning, involved the resources and expertise of 10 nations. Seven ships and eight aircraft took part, with more than 200 participants with scientific interests. Once again the NASA CV-990 aircraft was utilised over a period of one month, between June 2nd and July 2nd in 1984, with eight coordinated data flights to enable collection of measurements of sea ice, open ocean, land ice, and atmospheric processes at the surface. Other coordinated flights were conducted with the US Naval Research Laboratory’s (NRL) P-3, the NOAA P-3, a Canadian CV 580, and the French B-17. All flights were scheduled to coincide with overpasses of the NIMBUS-7 satellite, equipped with a scanning multichannel microwave radiometer (SMMR). Such large amounts of data were collected that very few analyses have been completed or results published yet. Otherwise the bulk of the results from ice surface and oceanographic experiments have been published in a recent special edition of the Journal of Geophysical Research (Vol. 92, No. C7, 1987).
1.3.2 MIZEX '84 and the scope of this study

This thesis represents work completed during a study to analyse data collected during MIZEX '84. The investigation as such only forms a small part of MIZEX as a whole, but to a large extent is encompassed by MIZEX East objectives (Johannessen, 1987). Figure 1.3 shows the main areas involved in MIZEX '84, and those dealt with in this study. The ice edge located on the map indicates the mean position of the outer limit of the MIZ, although the strong polar front and velocity shear caused by the southwards flowing East Greenland current favoured development of many complex ice features. These factors in combination with intense atmospheric frontogenesis caused rapid changes in ice form and concentration exhibited by the MIZ.

The timing of the experiment (mid June to late July) was chosen to yield as much information about the transition to summer conditions in the MIZ. Many cycles of melting and refreezing of the upper surfaces of the ice typically occur in the early part of this period, before typical summer conditions prevail. This was intended to provide a good range of conditions for micro-scale measurements of ice properties and thermodynamics, and for microwave remote sensing studies of the upper ice surface.

The most relevant aspect of the 1984 experiment was the use of a number of different aircraft carrying different sensors, enabling sequential observations of many highly variable ice phenomena during all weather conditions. The overall strategy involved remote sensing observations in several tiers, from satellite overpasses down to low level airborne flights. These multi-sensor observations were designed to overlap as much as possible to enable a synergistic use of data sets and cross-correlations between results of data analysis.

During MIZEX '84 the NASA CV-990 Airborne Laboratory (in Plate 1.1) was utilised in various missions. As part of the package of remote sensing equipment an updated version of the RAL radar altimeter was included onboard. The full suite of sensors was as follows;

1. RAL short pulse radar altimeter at 13.81 GHz, similar in specifications to the ERS-1 radar altimeter.

2. 19.35 GHz Imaging passive microwave radiometer, of specifications similar to the Nimbus 5 Electronically Scanning Microwave Radiometer (ESMR).

3. Fixed-beam microwave radiometers at 21 and 37 GHz.

4. Passive microwave imager at 92 GHz.

5. Two KS-87B Cartographic large-format cameras (5" film format).

6. Thermal infrared radiometer operating at 10.7 µm wavelength.
Figure 1.3 MIZEX '84 Operations plan, showing areas covered by shipborne observations and aerial remote sensing flights.

Plate 1.1 The NASA Convair 990 Airborne Laboratory during flight preparations.
The purpose of CV 990 flights was to obtain data to understand better the physical ocean/ice atmosphere interactions occurring in the marginal ice zone (MIZ), using a combination of the various sensor data, shipborne, and ‘on ice’ experiments. Fortunately, mission time was extended and an additional flight was planned over Nordaustlandet, Svalbard, to obtain a section of data over terrestrial ice.

Importantly, inclusion of the RAL radar altimeter on the NASA CV 990 aircraft enabled altimetric data to be collected over a variety of polar ice surfaces. Never before has a satellite-borne altimeter had a number of sources of supporting data sets, or data collected simultaneously from the same area on the ground. The data set acquired using the RAL altimeter is therefore unique in this respect, and gives an opportunity to investigate the mechanisms behind altimetric returns in different circumstances.

1.4 Radar altimetric studies of polar ice: an overview

The work completed by myself and presented in this thesis forms an integral part of a larger framework of several studies undertaken by a consortium of academic institutions, including the Scott Polar Research Institute (SPRI). The aim was to make detailed investigations into how a radar altimeter functions over a variety of different surfaces. This particular study represents a piece of work referring specifically to altimeter response over terrestrial ice and sea ice surfaces. As such it forms part of a study initiated in 1984 to investigate RAL radar altimeter response in polar regions, and reported by McIntyre et al. (1986). For a more comprehensive background to radar altimetry, the reader is referred to Rapley et al. (1983); Rapley et al. (1985); and McIntyre et al. (1986).

Data analysed were obtained during two different flights of the RAL radar altimeter. These are supported by field data acquired during MIZEX '84 and a number of other data sets. This thesis discusses the reduction and interpretation of these data with particular attention to the extraction of geophysical information therefrom. The developments to date in techniques of microwave altimetry are reviewed, and are updated with newer applications of this type of remotely sensed data using examples from land and sea ice. A number of limitations to this piece of research are recognised, being imposed by the instrument used, the data set collected, and the restricted seasonal conditions of the ice. Data were gathered on experimental flights of an essentially new instrument, and so instrumentation, and as a consequence the data, suffered from a number of teething problems. Data used here are selected for study using the criterion of minimum instrumental problems and data errors. Thus all results presented in this thesis are from sections where errors are minimal and where corrections (when possible) have been applied. It should also be noted that no attempt has been made to extrapolate these data to conditions which may exist in winter, early spring or autumn. Such extensions
of these results should be avoided unless surface data collection and experimental validation are possible at these times of the year.

The work undertaken in this thesis falls into four main categories:

(i) A review of the principles of altimetry, along with a technical description of the equipment used to gather data, and an experimental overview.

(ii) A theoretical background of electromagnetic interaction with snow and ice at GHz frequencies.

(iii) A review of the performance of an altimeter over sea ice and terrestrial ice.

(iv) An investigation of waveform characteristics and the geophysical information which may be extracted from altimetric data.

Part of the first group forms this Chapter, and a technical discussion and description of data processing techniques follow in Chapter 2. The three main areas of research make up Chapters 3 to 6. Chapter 3 is concerned with the electromagnetic properties of snow and ice at the relevant frequency (ie. 13 GHz). Interaction of microwave energy with these surfaces is treated in Chapter 4 using theoretical and empirical scattering models. Chapters 5 and 6 deal with specific aspects of altimeter operation over sea ice and terrestrial ice surfaces using examples from airborne altimetric data. The main characteristics are identified, and data products which may be extracted regarding properties of the surface are investigated. Finally Chapter 7 sets out the conclusions and main observations of the thesis, and makes suggestions for further work.
CHAPTER 2
SYSTEMS HARDWARE, DATA RECORDING, AND PRE-PROCESSING

2.1 The RAL altimeter

2.1.1 The RAL altimeter: an overview

The Rutherford Appleton Laboratory (RAL) radar altimeter is a high-resolution pulsed radar which operates at a frequency of 13.81 GHz ($\lambda = 2.2$ cm), and is intended for aircraft installation. It uses a pulse compression scheme based on a pair of matched surface-acoustic wave filters to provide an effective pulse width of 4 ns (McIntyre et al., 1986). Returned pulse waveforms are digitized with a sampling interval of 3.3 ns, and together with the two-way delay time and other relevant parameters are recorded by an HP1000 computer. Data are stored on magnetic tapes for subsequent processing and analysis.

High range resolution is achieved by using compressed pulses (ie. of short duration): the scheme used is similar to that used on the GEOS-3 satellite altimeter. Other satellite altimeters, for example; Seasat, Geosat, and the future ERS-1, use a different pulse compression technique called the 'full-deramp' method, which necessitates real-time tracking. The advantage of the RAL altimeter is is that no on-line tracking is required and the number of sampling gates (or range bins) may easily be varied up to a total delay window of 1 $\mu$s in width. This allows the altimeter to record consecutive pulse returns from surfaces with height variations of up to 150 m. Due to the limited received bandwidth of the receiver, the system point target response is of longer duration than transmitted pulses. The effective 3 dB duration is almost a factor of 2 longer than a compressed pulse, at 7.3 ns.

Three different antennas may be used on the altimeter, which are mounted on a sled for positioning in the underside of the aircraft fuselage. For the purposes of the MIZEX '84 campaign, transmission of pulses may be directed by the computer via one of two horns, or a paraboloid dish. Each horn was fixed and directed with its boresight axis pointing at nadir, while the parabolic dish was manoeuvrable and could be scanned in a cone about the vertical. Data discussed in this thesis were acquired by transmission and reception of pulses using one horn antenna only, since this technique enabled comparable like-polarised pulses to be analysed from different regions.

The main characteristics of the radar relevant to this study are given in Table 2.1. A full
block diagram of the radio frequency section is shown in McIntyre et al. (1986) along with a
detailed description of the main components of the instrument. The following section deals
briefly with several important aspects of system hardware.

Table 2.1  Nominal system specifications of the RAL altimeter (after McIntyre et al.,
1986).

<p>| | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Frequency</td>
<td>13.81 GHz</td>
</tr>
<tr>
<td>Travelling Wave Tube (TWT) Power</td>
<td>20.0W</td>
</tr>
<tr>
<td>Pulse repetition frequency (PRF)</td>
<td>100Hz</td>
</tr>
</tbody>
</table>

Pulse Compression System:
- Transmitted pulse length | 320 ns
- Bandwidth               | 320 MHz
- Compressed pulse width  | 4 ns
- Compression ratio       | 80 (19 dB)

Pulse Digitization:
- Sampling frequency       | 300 MHz
- Resolution               | 5 bits + sign bit
- Input bandwidth          | 100 MHz
- Total number of delay channels | 1024

Antennas: 3 dB Beamwidths (degrees)
<table>
<thead>
<tr>
<th></th>
<th>E-plane</th>
<th>H-plane</th>
<th>Gain (±0.35 dB)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horn (X)</td>
<td>8.9</td>
<td>13.1</td>
<td>23.3</td>
</tr>
<tr>
<td>Horn (Y)</td>
<td>9.1</td>
<td>13.2</td>
<td>23.4</td>
</tr>
<tr>
<td>Parabolic Dish</td>
<td>3.1</td>
<td>2.9</td>
<td>33.05</td>
</tr>
</tbody>
</table>

2.1.2 Horn antenna characteristics

The two horn antennas are identical, each having approximately elliptical beam patterns
 corresponding to beam widths between 3 dB points of 13 and 9° along the major and minor
 axes respectively. Ulander (1985) calculates the ‘effective’ beam width of the horn antennas
 by modelling the antenna beam pattern with a Gaussian function. He derives the following
 expression to calculate the effective circular 3 dB half-beam width;

\[
\frac{1}{\sin^2 \theta} = \frac{1}{2} \left( \frac{1}{\sin^2 \phi_E} + \frac{1}{\sin^2 \phi_H} \right)
\]

\{2.1\}

where \( \phi_H \) and \( \phi_E \) are the half-beam widths in the two planes of symmetry. Substituting
values given in Table 2.1 gives 5.25° for \( \theta \). This corresponds to a full 3 dB beam width of
10.5°, which is only slightly lower than the arithmetic mean of the two values for the angles
in the two planes of symmetry. This is a useful concept in later sections when during analysis of altimetric data it becomes necessary to assume that the target illumination is circular and statistically homogeneous.

When installed in the aircraft the antennas point at nadir with orthogonal polarisations. X-horn is oriented to transmit an E-field parallel to the centreline of the aircraft fuselage, and the Y-horn is fixed orthogonal to it. The polarisation of the X and Y-horns is potentially misleading, and does not refer to horizontal (H) and vertical (V) polarisations. Both are in fact vertical in the normal flight configuration.

A special arrangement enables each of the horns to be used for transmission or reception. The main use of this facility is to allow cross-polarisation measurements in which power is transmitted from the X-horn and received by the Y-horn (or vice versa). Antenna selection is initially made manually but is performed under computer control during operation.

2.1.3 Pulse transmission and control logic

Transmitted signals may be produced at varying intervals. The pulse repetition frequency (PRF) could be varied with pulse separation intervals ranging between 1 and 15 ms in 1 ms intervals. In practice, higher PRF's were not used because the resulting data rate was too high for the software to handle (Birks; personal communication, 1984). All observations in this study were made with a PRF of 100 Hz.

A chain of counters forms the important part of the radar control logic. The principal output of this unit is a transmit trigger signal, which indicates that the system is armed and ready for pulse transmission. This signal is repeated at the PRF. Further control signals are produced which select the appropriate antenna, and arm the receiver and a time interval counter (at an appropriate point in the transmit/receive cycle) in fixed relation to the transmit trigger.

2.1.4 Pulse reception and signal digitizing

Signals can be received and recorded in a number of modes using different antenna configurations and receiver gains. In order to enable receiver gain control three programmable IF attenuators were included in different signal paths, so that the total attenuation could be alternated between two distinct values by simply switching received signals from one path to the other. One such mode was used on the 28th June flight over Nordaustlandet, Svalbard, when the paraboloid antenna was utilised as a 'look-ahead' beam with a nadir-directed horn antenna. Attenuator settings for each flight day are shown in Table 2.2. A calibration path is built into the instrument which can divert a fraction of the transmitted signal through a
waveguide switch connected between the antenna and receiver systems. Transmitted power could thereby be monitored periodically using this system.

Detector output is routed to a Biomation model 6500 Waveform Digitizer. This unit was capable of sampling and digitizing the signal. The digitizer measures the shape of the echo, and records the envelope of received power.

Table 2.2  Table of attenuator settings and modes of operation for data analysed.

<table>
<thead>
<tr>
<th>Date</th>
<th>Attenuator Settings</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Odd</td>
<td>Even</td>
</tr>
<tr>
<td>June 28th</td>
<td>7 dB</td>
<td>15 dB</td>
</tr>
<tr>
<td>June 28th</td>
<td>10 dB</td>
<td>6 dB</td>
</tr>
</tbody>
</table>

Note. For land ice modes;  
Odd channel=Horn  
Even channel=Dish  

For alternating sea ice modes;  
Both channels=Horn  
Odd channel=XX  
Even channel=XY

2.2 Data recording

The hardware controlling the radar consists of a Hewlett Packard HP100F minicomputer. Peripherals include a graphics VDU, thermal printer, magnetic tape drive and a 20 Megabyte disc drive. Normally the computer also controls data collection and the control software, and is interfaced with a waveform recording unit to synchronise switching control signals, to set up different recording modes, and to select different antennas. Several aspects of data recording are discussed below.

2.2.1 Signal digitizing

Sampling and digitizing of the detector output is possible with 6-bit resolution (5 bits + sign). The instrument provides an internal sampling clock which enables signals to be sampled and digitized at rates of up to 500 MHz, and also provides the option of using an external clock source. The RAL altimeter uses an external sampling clock at a fixed frequency of 300 MHz in order that trigger signals could be synchronised with the sampling clock (McIntyre et al., 1986). The input bandwidth of the unit is specified as 100 MHz. This means that the point
target response is broadened by the digitizer, but that a sampling frequency higher than 300 MHz is not necessary.

The Biomation waveform digitizer has a 0.5 Volt input range, and 0 to 32 digitizing levels. One Biomation amplitude unit (bau) thus corresponds to 15.6 millivolts of returned power. Since signal amplitudes are discussed frequently in subsequent chapters, units of bau are preferred to Volts. For this reason all representations of return waveforms have a maximum limit of 32 bau. In addition there is a background d.c. voltage from the detector which is present irrespective of the level of the detected pulse or the IF attenuator setting. It introduces an offset of generally 4.2 bau at the Biomation output, and this background level may be observed in all recorded signals prior to the main return waveform. Before the recorded data can be interpreted as detected signal level, the offset has to be subtracted.

Instrument memory is sufficient to store 1024 consecutive samples. Each individual range bin has a resolution of 3.33 ns, and so the total sample or ‘delay window’ may be extended to 3410 ns. In practice full capacity for storage was never used; sea ice altimetry used only 100 samples, and land ice used 196.

To record data, the digitizer must be armed and ready to receive information. Upon transmission of a pulse, a signal is sent to arm the system. Upon receipt of this signal it stores samples continuously, until receipt of another ‘trigger’ pulse, and thereafter for a fixed delay. Once this delay has elapsed, no more samples are stored, and the last 1024 samples contain information which may be recorded, depending upon the size of the delay window. Each recorded pulse is allocated a sequence identifier upon triggering of the digitizer, which allows it to be time-tagged even when not all pulses are received or cause triggering.

2.2.2 Trigger options

Two trigger modes are possible with the Biomation digitizer. These differ in the source of the trigger signal to the recording unit.

2.2.2.1 Delay mode

The delay trigger is simply a system whereby a trigger signal occurs at a fixed point in the time base determined by the computer. Although in principle this would allow a rudimentary tracking loop to be implemented, the software maintains a fixed delay. The delay value, preprogrammed by an operator, is determined according to the aircraft operating height. It is periodically changed so that the tracking window spans the expected range of delay of returned echoes. For observations over sea ice this system was utilised for waveform recording. It has, however, since been proved to have inherent problems, the main one being that the aircraft cannot maintain a perfect fixed altitude, instead flying with oscillatory vertical movements.
Consequently the positioning of recorded waveforms within the delay window is variable, wandering towards the early section when the aircraft reduces altitude, and towards the latter part with increases in altitude. A common result is that waveforms become truncated by either limit of the recording capacity. Such sections of sea ice data have had to be rejected in favour of sections where the fixed delay was changed frequently to maintain central positioning of recorded signals within the delay window.

### 2.2.2.2 Video mode

This mode derives the trigger signal from the received echo itself. When the output of the detector exceeds a fixed reference level, the trigger synchronisation logic produces a trigger pulse synchronised with the next cycle of the 300 MHz sampling clock, enabling detector output to be recorded. This mode known as video trigger is employed with altimetry over land ice, as changes in terrain clearance are continuous and sometimes rapid. Although this method is generally successful, one problem is encountered with echoes whose amplitude does not exceed the trigger threshold. Such instances are common over surfaces with low backscatter coefficients, and result in intermittent triggering and data gaps.

### 2.2.3 Real-time data handling

The main real-time tasks involve ensuring that every digitized echo is correctly associated with a timer value, and that recorded pulses are stored efficiently on magnetic tape. Digitized data are stored on tape as a ‘minor frame’. Each minor frame is assigned the most recent time read out from the computer clock, and the value of the Biomation interface counter (indicating which antenna is used for transmit and receive). The computer control software assembles a predefined number of echoes (between 90 and 1000) into a larger group known as a ‘major frame’. The size of such groups is determined by the size of each minor frame. Between major frames there is a gap of about one second, during which period echoes are not recorded, to enable all control settings and the position of the parabolic antenna to be interrogated and recorded.

### 2.2.4 Data quality indicators

There are a number of ways of distinguishing sections where faulty data occurs. In the first two instances data may not be recorded as a result of certain problems, and data quality here refers to the proportion of transmitted pulses that give rise to echoes that are received and recorded.

Firstly, particularly when a large number of delay samples are being recorded by the Biomation digitizer, the computer is not always fast enough to process and record every return
in real-time. This occurs in places with the Nordaustlandet data set, and causes overflow in the Biomation interface counter (BIC). Such records are distinguishable by a BIC number which is not sensible (usually negative), instead of a normal cumulative clock value. Additionally, instead of measuring the delay correctly in these instances, spurious values can be recorded, by measuring until reception of the next pulse. These records must therefore be ignored during data analysis.

Secondly, as explained, during video trigger operation some echoes may be too weak to cross the trigger threshold, and will thus not be detected. The number lost will be dependent upon the signal-to-noise ratio. The proportion of echoes recorded can be deduced from the time taken to record each major frame (available in the data listing on magnetic tape), and also a fault count. Sections of data where the fault count is particularly high are avoided in subsequent analysis, where possible.

The third instance is associated with data which are recorded by the computer, but which are flagged as spurious data. These binary flags, assigned values as each minor frame is recorded, may be checked before analysis of each record. In this way data may be either accepted or rejected. There are 6 important flags in all, of which 4 are used routinely during data analysis.

The first flag should always be assigned a binary 1, to signify that the horn antennas are being used in a vertical mode rather than the steerable paraboloid antenna. The second is a synchronization flag which should always be 0 if the control logic has performed correctly. Third is an address error flag which indicates problems of data storage (if set to 1), and fourth is a duplicate data flag (indicating bad data if set to 1). Only the first of these flags should be set intentionally, and so all records where the remaining flags are set to 1 are rejected.

2.2.5 The airborne digital data acquisition system

On board the CV-990 was an Airborne Digital Data Acquisition System (ADDAS) dedicated to in-flight housekeeping of various recorded parameters. The system samples at a frequency of 1 Hz, recording all the parameters shown in Table 2.3, except the Inertial Navigation System (INS) output, which is updated every 2 seconds.

<table>
<thead>
<tr>
<th>Parameter</th>
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</tr>
</thead>
<tbody>
<tr>
<td>Latitude</td>
<td>Pressure Altitude</td>
</tr>
<tr>
<td>Longitude</td>
<td>Radar Altitude</td>
</tr>
<tr>
<td>Ground Speed</td>
<td>Pitch</td>
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<tr>
<td>True Heading</td>
<td>Roll</td>
</tr>
<tr>
<td>True Air Speed</td>
<td>Wind Direction</td>
</tr>
<tr>
<td>Static Air Temperature</td>
<td>Wind Speed</td>
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Table 2.3 The major parameters recorded by the ADDAS onto magnetic tape.
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Table 2.3 The major parameters recorded by the ADDAS onto magnetic tape.

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<tr>
<td><strong>Static Air Temperature</strong></td>
<td><strong>Wind Speed</strong></td>
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2.3 Pre-processing stages

A number of tasks must be completed before altimeter waveforms can be analysed or understood. This stage in the investigation of the raw digital information is regarded as pre-processing of the data. However, before this stage may be undertaken the statistical properties of the recorded data must be investigated.

2.3.1 Statistical properties of RAL altimeter waveforms

The RAL instrument is a monostatic radar (ie. transmits and receives from the same antenna) and transmits coherent pulses. Unfortunately single pulse waveforms returned to the altimeter have no validity when used independently, since they display rapid fluctuations and a 'noise-like' appearance. This phenomenon is known as 'fading' of the signal, and results from coherent addition of backscattered energies from randomly distributed scattering elements upon the surface. Since this is a fundamental limitation when considering radar returns from spatially inhomogenous surfaces such as sea ice and terrestrial ice, the effects of fading must be investigated and an optimal sample size devised for the averaging of raw waveforms over all surface types.

In the context of the RAL altimeter the number of point scatterers within the footprint must be assumed large. Their relative locations must be random and no single scattering element should dominate the backscattered signal. In such a system fading is reasonably well understood (Ulaby et al., 1982). At any single point in time, after a radar pulse has been backscattered, the signal received by the antenna is a phasor sum of the contributions from all the point scatterers. This sum varies in time and space depending upon the scattering characteristics of the point scatterers. Doppler frequency shifting of the returns results in marked fluctuations in signal amplitude.

Upon examination of the probability density functions (pdf's) derived from RAL altimeter data, it is confirmed that for the RAL square-law detector the pdf which characterises fading is of exponential form (Francis et al., 1982). An exponential distribution is a one parameter pdf where;

\[ p(P) = \frac{1}{\alpha} \exp \left( -\frac{P}{\alpha} \right) \]  

where \( P \geq 0 \), \( \alpha \) characterises the distribution, and \( P \) is the sampled power. McIntyre et al. (1985) show that the maximum likelihood estimator for \( \alpha \) is simply the mean power \( \bar{P} \), and characterises the shape of a distribution thus;

\[ p(P) = \frac{1}{\bar{P}} \exp \left( -\frac{P}{P} \right) \]  

where \( \bar{P} \) is the mean of a sample of \( P \). In this case the standard deviation is equal to the mean, this being quite different from a normal or Gaussian distribution where the mean and
standard deviation are unrelated. In conclusion, the best way of deriving a good estimate of the mean pulse return is by summing a sequence of pulse waveforms and calculating the arithmetic mean of the powers in each of the range bins.

There is a strict tradeoff between, on the one hand maximising sample size utilised for averaging to increase our confidence in estimates of mean power, and on the other hand minimising distance travelled between contiguous averaged waveforms, to increase spatial resolution. The effects of sample size upon the individual estimates of power values in mean pulse returns must therefore be examined. Pdf’s derived from RAL data are displayed in Figure 2.1 for different sum totals showing that power levels are, as expected, approximately exponentially distributed. Confidence limits may be based on the fact that the sum of \( N \) samples from an exponential distribution is chi-squared distributed with \( 2n \) degrees of freedom. Since the standard deviation (\( \sigma \)) for the average is given by \( \overline{P}/\sqrt{N} \) the improvement of the estimate of mean power in an average waveform is proportional to \( \sqrt{N} \). The 95% confidence limits for four pdf sample sizes are displayed in Table 2.4 below:

<table>
<thead>
<tr>
<th>( N )</th>
<th>Integration Distance ( (m) )</th>
<th>Integration Time ( (s) )</th>
<th>95% Confidence limits for ( 2n ) Degrees of Freedom</th>
</tr>
</thead>
<tbody>
<tr>
<td>10</td>
<td>40</td>
<td>0.2</td>
<td>( 0.59\overline{P} &lt; \text{mean} &lt; 2.10\overline{P} )</td>
</tr>
<tr>
<td>20</td>
<td>80</td>
<td>0.4</td>
<td>( 0.67\overline{P} &lt; \text{mean} &lt; 1.64\overline{P} )</td>
</tr>
<tr>
<td>50</td>
<td>200</td>
<td>1.0</td>
<td>( 0.77\overline{P} &lt; \text{mean} &lt; 1.35\overline{P} )</td>
</tr>
<tr>
<td>100</td>
<td>400</td>
<td>2.0</td>
<td>( 0.81\overline{P} &lt; \text{mean} &lt; 1.20\overline{P} )</td>
</tr>
</tbody>
</table>

A standard deviation of 10% of the mean power thus requires an sample size of over 100 single pulses, yet the integration distance on the ground is almost 0.5 km. To choose the number of of pulses used to construct mean waveforms, one must therefore consider the tradeoff between improved precision in power estimates, and the spatial smoothing caused by large values of \( N \). For the purpose of this study, an average in the range of 50 to 100 was chosen. The justification for this decision lies with the fact that the sea ice/water mixtures and terrestrial ice surfaces investigated in later chapters are of limited spatial stationarity. It is necessary to accept increased variance due to inadequate averaging of fading effects, in order that the profiled terrain does not change significantly. A value of \( N = 50 \) or more, appears to be reasonable, since the 95% confidence bands in Table 2.4 correspond to a range of 2.5 dB or less. Such uncertainty in a single estimate of mean power, for example, must be taken into account when contiguous mean pulse waveforms are compared or geophysical parameters are
extracted. The 2.5 dB range, although large, may still enable the distinction between: (a) bare ice and dry snow, which have a possible relative difference in Fresnel reflection coefficient of up to 7 dB, (b) wet snow and dry snow, which have a possible relative difference of up to 11 dB, and (c) ice and water with a difference of around 12 dB. Thus algorithms for the extraction of geophysical information from mean waveforms may still be derived.

Figure 2.1 Histograms of power in a single range bin in waveforms produced by averaging 10, 20, 50, and 100 consecutive return pulses. Note that no power is observed below 4.2 bau due to the d.c. offset discussed in section 2.2.1.

2.3.2 Pulse-to-pulse decorrelation

It is clear from the previous section that there is an optimal sample size to achieve a good estimate of mean pulse waveforms. Before making measurements or deriving geophysical parameters from the shape of the mean pulse return it is necessary to ensure that statistically independent or 'decorrelated' pulses (Walsh, 1982) are averaged. The accuracy of the resultant measurements rest upon the number of such successive pulses included in the average. If pulse-to-pulse correlations exist within averaged pulse waveforms then more pulses than usual will be required in order to achieve the confidence levels in Table 2.4.

The variation between successive received pulses results from the motion of the radar over the surface. At any instant, as previously explained, the electric field sampled by the antenna is the phasor sum of the fields scattered from many individual point elements within the illuminated footprint. With movement the relative phase of the contributions from these elements will change. Providing that this change is sufficient from one pulse to the next then successive individual pulses will be decorrelated. For the RAL altimeter McIntyre et al.
(1986) calculate that successive returns from the pulse-limited footprint (PLF) are statistically independent if the interval between pulses exceeds \(1/\Delta f\) (where \(\Delta f\) is the Doppler frequency shift). \(\Delta f\) may be calculated using the following equation for the RAL altimeter:

\[
\Delta f = 3.464 \frac{vf_H}{\lambda H}
\]  

where \(v\) is the aircraft velocity, \(\lambda\) is the wavelength, \(H\) is the aircraft height, and 3.464 is a factor determined by the shape of the Doppler spectrum. The radius of the pulse-limited footprint \((r)\) on a flat-diffuse surface is

\[
r = \sqrt{c \tau H}
\]  

where \(c\) is the velocity of light, and \(\tau\) is the pulse width. For the RAL altimeter; \(v = 200\text{ms}^{-1}\), \(H = 10\text{km}\), and \(\lambda = 2.2\text{cm}\). Using equations 2.1 and 2.2 a value of 474 Hz is derived for the Doppler frequency shift, corresponding with an interval of 2.1 ms. Thus, providing that the interval between successive RAL pulses exceeds this value of \(1/\Delta f\), then individual successive returns are uncorrelated. The maximum frequency used during the flights analysed here was 100 Hz (ie. a 10 ms interval), and the normal PRF was 15 ms. The criterion for pulse separation is therefore satisfied for returns from within the pulse-limited footprint (PLF). Since returns originating from outside the PLF will have increased delays, so too will they have a correspondingly increased Doppler spread. If samples of successive pulses from the PLF with smaller delay times are uncorrelated then it follows that those with higher delays will also be.

2.3.3 Waveform tracking and averaging

This section is concerned with algorithms to obtain mean pulse waveforms, thus enabling the extraction of geophysical information. In order to reduce the statistical fluctuations caused by fading it is essential to average individual pulse returns. Various schemes have been suggested, to enable individual pulses to be correctly superimposed (McIntyre et al., 1986; and Ulander, 1987), with differing degrees of success. Tracking is necessary to take into account changes in range relative to the target, variations in aircraft motion, and changes in the shape of individual pulse returns. This is because there is insufficient memory to store the return signal from the moment of transmission.
2.3.3.1 Threshold tracking

The threshold tracker is the most simple tracking technique which detects pulse returns with a threshold detector (Skolnik, 1970) and measures the time delay to that point. Individual pulse returns are aligned relative to the track point before averaging. Corresponding range gate detector counts are summed and averaged accordingly, to give an estimate of the mean pulse waveform. The advantages of threshold tracking with RAL altimetry data are that the delay time for the pulse to be received is not necessary, and so waveforms could be averaged for which there were no reasonable delays. It enables precise summation of pulses over flat surfaces, and through reduced pulse blurring gives an accurate determination of range to the target.

It is clear that this form of tracker is not suitable in certain circumstances, since single pulses suffer from considerable fluctuations in power due to fading. Figure 2.2 shows a sequence of individual recorded waveforms from over the ocean, which demonstrates this phenomenon.

![Figure 2.2](image)

Figure 2.2 Series of 8 consecutive raw pulse waveforms reading across the page. The waveforms were recorded over open ocean and display typical fading characteristics, such as the rapid fluctuations in power.

The problem arises in locating the first high amplitude peak of each waveform and superimposing all peaks over an arbitrary limit. The resulting summing process adds all energy contained in the early peaks of returns coherently. Ulander (1987) identifies three artefacts of the tracking process.

(a) The mean waveform value at the track point is highly biased, and appears as a spike.

(b) Mean pulse waveforms have distorted leading edges, whose gradient is too steep.

(c) All high amplitude returns at later delays are smoothed because of their random offset.
in terms of time delay relative to the track point.

It is noted by Ulander that the effects of the threshold tracker are reduced by decreasing the threshold level. A perfect mean return can be recreated in the limit where the threshold is 0, but this is not realistic since noise fluctuations would be tracked. In the case of the RAL altimeter the tracker threshold must always be larger than the d.c. offset of 4.2 bau, and high enough above that level to escape spurious tracking on noise. The result of such artefacts is that in cases where information is desired from the initial portion of mean pulses, it is impossible to derive estimates which have geophysical relevance.

2.3.3.2 \( \alpha - \beta \) tracking

An \( \alpha - \beta \) tracking system was first used on Seasat-A radar altimeter waveforms (MacArthur, 1978), and a modified version is planned for the ERS-1 mission (Levrini et al., 1984). The essence of the tracker is that it calculates the range \( h(n) \) and range rate \( \dot{h}(n) \) from the history of the signals, and predicts the range \( h_p(n + 1) \) at intervals of time \( \Delta t \) for each successive pulse return. The following equations are used in a recursive fashion:

\[
\begin{align*}
  h(n) &= h_p(n) + \alpha \Delta h(n), \\
  \dot{h}(n) &= \dot{h}(n-1) + \beta \Delta h(n)/\Delta t, \\
  h_p(n + 1) &= h(n) + \dot{h}(n)\Delta t 
\end{align*}
\]

where \( \Delta h(n) \) is the range error, and \( \alpha \) and \( \beta \) are tracker constants. A suitable choice of the constants will enable calculation of a predicted range and reliable tracking of the surface. The parameters \( \alpha \) and \( \beta \) are interlinked, if noise in tracking is to be suppressed efficiently. Benedict and Bordner (1962) choose \( \beta = \alpha^2/(2 - \alpha) \) for optimal noise suppression, while a critically damped tracker with \( \alpha = 2\sqrt{\beta} - \beta \) is favoured by Cadzow (1973), for its improved damping. This is undoubtedly a suitable system for the tracking and averaging of pulses from a homogeneous and slowly varying surface such as an ocean. It gives rise to undistorted mean pulse waveforms, which is important if theoretical pulse shapes are to be matched, for extraction of information regarding the ocean surface.

The \( \alpha - \beta \) tracker on the Seasat radar altimeter had four modes, employing different tracking parameters. Mode 1 used a critically damped selection, while modes 2 and 3 were close to the optimum. The remaining mode enabled parameters to be selected from the ground. Unfortunately, since the mission lifetime was so short, mode 1 was used almost all the time (Rapley et al., 1983).

The drawback of this scheme is that the tracker is not agile enough over rapidly varying topographic surfaces. A more agile tracker could have been used by choosing different values for \( \alpha \) (Ulander, 1987), providing that increased noise in predicted range is acceptable. In a wider
context the system is not robust enough when rapidly changing waveforms are encountered. The height error estimate $\Delta h$ is only reliable when the shape and intensity of pulse returns are uniform. If either vary quickly, a bias results, causing pulse blurring, additional range jitter, and in extreme cases loss of lock (i.e. loss of the pulse outside the scope of the range window). The problems of employing this type of tracking scheme are far greater for an airborne altimeter because the vertical motion contains rapidly varying components. The short-term motion of the aircraft has been shown by McIntyre et al. (1986) to be quasi-periodic with a period of approximately 20 seconds.

An additional limitation, which is inherent to the $\alpha - \beta$ concept, is that it relies upon consecutive data to update the tracking loop. Sudden data gaps will cause loss of lock, and so the existence of inter-frame gaps in data records makes this type of tracker unsuitable.

### 2.3.3.3 Interpolation tracking

The problems associated with tracking and averaging waveforms identified in the previous sections prompted attempts to develop a more robust algorithm (Ulander, 1987). The scheme is based on concepts similar to both the threshold and $\alpha - \beta$ trackers, and its aim is to provide unbiased average pulse returns in the presence of statistical fluctuations with minimal distortions.

![Interpolation Tracking Principle](image)

**Figure 2.3** Schematic diagram illustrating the interpolating procedure. A least squares fitting process enables calculation of the range rate from the gradient of the regression line, and the standard deviation of range delay $\sigma_t$ is equal to the rms roughness of the surface.

The interpolation tracker is different from the $\alpha - \beta$ algorithm in the way it estimates range
and range rate. It first obtains a sample of individual data records, and fits a linear regression to the group of typically 50 to 100 pulse delays. Figure 2.3 shows the principle of fitting a smooth line through the data points enabling prediction of the delay to the mean surface. Individual recorded delay times are corrected for their offset from the regression line, enabling pulse returns to be correctly superimposed and averaged. This is based on the assumption that transmitted pulses are backscattered from the ice surface at nadir. Figure 2.3 shows that the rate of change of delay caused by the regional slope of the surface is given by the gradient of the regression line. In addition the standard deviation of the delays about the line may be used to estimate the rms roughness of the surface being tracked.

The tracker must predict the range and filter the range rate using this method, with the assumption that the variation is linear or constant. Thus the number of pulse returns used in the averaging process has to be considered in order to maintain a constant range rate. Ideally, the larger the sample size the higher the confidence level for the least-squares range rate estimate. Dotted lines in Figure 2.3 indicate errors caused by the least-squares fitting process. These cause pulse ‘blurring’ by spurious corrections of the delay times and incorrect superimposition of individual pulses during averaging. It is evident that a reduction in the number of data points increases the inaccuracy of the fitting process. Ulander (1987) shows that the the standard deviation of pulse spreading in range units ($\sigma_z$) is

$$\sigma_z = 2\sqrt{3} \frac{\sigma_t}{\sqrt{N}}$$

where $\sigma_t$ is the standard deviation of pulse delays, and $N$ is the number of pulses in the sample.

The performance of the interpolation tracker is investigated by Ulander (1987) for the RAL altimeter. As a measure of its performance the error in the estimate of the rms surface height is calculated, showing that $N$ needs to be more than 200 to keep errors to less than 0.1 m. The trade-off between spatial smoothing and measurement precision discussed in relation to fading is therefore also of relevance here. For a sample size of 100 pulses the error in rms roughness estimation is reduced to below 5.0% for surfaces with rms roughnesses of 2 m or more. Any reduction in sample size causes a corresponding exponential reduction in the accuracy of rms roughness estimation.

The algorithm has the advantage over an $\alpha - \beta$ tracker that it is not affected by the inter-frame gaps in RAL data storage, and although it was designed with the video trigger in mind it is also useful for delay trigger waveforms.
2.3.3.4 Trackers: general observations

The previous brief look at trackers shows that several schemes have been investigated for producing accurate estimates of mean pulse waveforms. The selection of the latter tracking algorithm was based upon a number of important factors. Firstly, there are a number of data considerations, such as the gaps between major frames, and the interleaving of pulses received in different antenna modes. Secondly surfaces are shown to have spatially varying scattering characteristics, requiring agile tracking algorithms over both land and sea ice. Finally, the interpolation tracker is tailored to the RAL altimeter data set incorporating a number of complementary algorithms to find and remove spurious data, caused by instrument malfunction and/or problems of mistriggering (discussed in section 2.2.4). These act as a data quality check, and include sorting routines which guard against spurious data which are geophysically invalid becoming incorporated in the analysis.

In the following sections the interpolation tracker is applied to both land ice and sea ice data sets. Testing of the tracker was not considered necessary since Ulander (1985), McIntyre et al. (1986), and Ulander (1987) have investigated its merits relative to other tracking algorithms. Range gate statistics have been calculated to show that estimates of power from interpolation-tracked mean waveforms have an exponential distribution (accounting for the 4.2 bau d.c. offset). This follows the pattern expected for a square-law detector from the theoretical analyses in subsection 2.3.1.

2.3.4 Range delay corrections and surface elevation measurements

Range may be measured accurately with the RAL altimeter with the values of delay recorded for each individual waveform. First, to enable these raw delays to be used, a number of corrections must be applied, since the interval of time recorded does not represent the absolute delay to the surface (ie. precise time between pulse transmission and reception).

A section of data are chosen when the altimeter is operating in video trigger, and where the aircraft crosses from land to a sea surface, to illustrate the errors in using the raw delay values. The raw uncorrected altitude data from the RAL altimeter records are plotted alongside both ADDAS radar altimeter and pressure altimeter records in Figure 2.4. Measurements over the flat invariant ocean surface indicate an inaccuracy of approximately 520 m when compared with the equivalent ADDAS radar altimeter records. Several corrections must be made therefore before delays can be used to represent an accurate measure of terrain clearance.

The real delay is a combination of the delay measured between the timer start trigger and the stop trigger produced by the video trigger signal \( T_{\text{rec}} \), and the following measured errors:
There is a fixed lag of 5 $\mu$s between transmit trigger and the timer start trigger.

A signal delay of 1.54 $\mu$s occurs within the transceiver.

The waveguide in the antenna system causes a two-way delay of 27 ns.

A trigger delay of 33.3 ns occurs within the Biomation.

The corrected delay $(t + \Delta t)$ in the video trigger mode is thus

$$t + \Delta t = T_{rec} + 3.466 \times 10^{-6},$$  \hspace{1cm} \{2.10\}

and the time correction in terms of range is normally 519.95 m. However, an extra correction must be applied when the altimeter operates in the delay trigger mode, due to the manual Biomation delay setting. This extra delay occurs between the Biomation stop trigger and pulse reception, and a range correction must be varied to correspond with changes in the setting used on each flight (depending upon the aircraft operating altitude). For the flight of June 30 discussed in chapter 5, the altimeter operated in delay trigger, and had a series of six changes in the Biomation delay setting. The corresponding delay corrections range from 2.2$\mu$s to 2.63$\mu$s (ie. altitude corrections of between 330 m and 395 m). These different corrections must be applied at intervals throughout the flight.

![Figure 2.4](image)

**Figure 2.4** Corresponding sections of RAL and ADDAS altimetry illustrating necessary corrections before accurate terrain clearance measurements or surface elevations may be calculated.

It is also evident from plots like Figure 2.4 that the ADDAS pressure altimeter suffers
from spatial changes in barometric pressure. During the period of time plotted the resulting bias is some 150 m. If, over terrestrial ice masses, accurate elevations are to be calculated, the RAL altimetry must first be corrected using the technique explained above, and second corrected to sea level with pressure altimeter records. Since biases are invariably present between ADDAS pressure and radar altimeter records a full history of their difference has had to be reconstructed for each of the flights. This is achieved by recording the bias at regular intervals throughout ADDAS data records. This enables interpolation of the bias correction for pressure altimeter records which are necessary to reconstruct surface elevations accurately.

2.3.5 Implications of an aircraft platform

It is important to consider in what ways the aircraft influences the operation of the RAL radar altimeter and the results obtained. The most important factors are the sensor-terrain range, and the motion and stability of the aircraft in flight.

2.3.5.1 Sensor-terrain range

A constant altitude was maintained of approximately 10 km during all flights from which data are analysed. Importantly, the terrain clearance only varies significantly over Nordaustlandet, where surface elevations exceed 700 m in places.

Major effects are caused by differences in the terrain clearance. Importantly, for specular surfaces such as calm seawater, backscattered power is inversely proportional to the square of the sensor-terrain range. For a perfect isotropic scatterer the contrasting level of backscattered power is inversely proportional to the sensor-terrain range to the third power. If the relative strengths of coherent and diffuse components in altimeter signals are calculated for a nominal altitude of 10 km, then the amplitude of the specular component is likely to be of the order of 10 times larger. This has important consequences both for the recording of return waveforms over surfaces with rapidly varying scattering characteristics, and in the analysis sections in later chapters.

With a 10° 3 dB full beam width for the horn antennas, at 10 km altitude the resultant nominal beam limited footprint (BLF) over sea ice and ocean surfaces is 1.75 km in diameter. The diameter of the pulse limited footprint (PLF) is calculated using equation 2.5. The pulse width from Table 2.1 is 4 ns, and the calculated footprint diameter $D_{PLF}$ at an altitude of 10 km of 219 m is plotted to scale in Figure 2.5 for comparison with the beam limited case. Over Nordaustlandet, Svalbard, during the flight on June 28, the maximum surface elevation encountered is 780 m (Dowdeswell; personal communication, 1986). In cases such as this, where terrain clearance is significantly reduced, the maximum change in $D_{PLF}$ is calculated as only 10 m, and so corrections are not considered necessary for the changes in illuminated
area over flat surfaces. Much larger changes in the area illuminated occur where the roughness amplitude exceeds the transmitted (compressed) pulse width. Importantly, a 4 ns pulse width is correspondent with 1.2 m vertical height. However, provided local surface undulations are less than \( c \tau \), 219 m may be regarded as the PLF diameter and the illuminated area assumed to be constant. The simultaneously illuminated footprint is a factor of \( \sqrt{2} \) larger than the PLF, giving a diameter of 309.7 m for flat non-sloping surfaces.

**Figure 2.5** Beam geometry for an aircraft altimeter operating at a nominal altitude of 10 km. The diagram shows how the illuminated footprint varies in size at critical points in time. \( N \) is the nadir point, \( c \tau \) is the pulse width, \( D_{PLF} \) is the pulse limited footprint diameter, \( D_I \) is the diameter of the simultaneously illuminated footprint, and \( D_{BLF} \) is the diameter of the beam limited footprint. Return pulse travel times are indicated.

In theory, for a flat non-sloping surface, the range of incidence angles encountered by transmitted pulses and recorded within digitized pulse waveforms, is determined by the duration of the range window. Although for a wide-beam, pulse limited altimeter such as the
RAL instrument, the footprint annulus expands as far as the 3 dB beam limits, no data are recorded beyond the footprint determined by the range window size. The diameter of this limit is calculated for the point in time when no further backscattered power is recorded by the waveform digitizer. At a terrain clearance of 10 km over sea ice the range window duration is 333 ns, and the range window footprint (RWF) is 1.42 km in diameter. The half-angle subtended at the circumference of the RWF is ultimately what determines the backscatter response characteristics of the instrument.

The above factors ultimately determine what the altimeter 'sees', and the amounts of information which it is possible to extract from altimeter data. Beam geometry and sensor-terrain range are observed to have important consequences, particularly for surfaces such as marginal sea ice which are spatially inhomogeneous and rapidly varying in scattering characteristics.

2.3.5.2 Platform instability

Aircraft oscillate in flight in the vertical direction, in a manner described as 'porpoising' or a 'Dutch roll'. This motion is incorporated in the delay times of the RAL altimetry data records, and also recorded by several of the ADDAS sensors.

Figure 2.6 Aircraft porpoising, as recorded by the ADDAS pressure altimeter during normal flight attitude over open ocean. The true airspeed is indicated by the upper line and right hand axis. The altitude readings show a number of components of vertical oscillation.
Figure 2.6 shows this motion, and analysis of these and other data gives a mean period of oscillation of 21 s and a standard deviation of range oscillation period of 2.7 s (McIntyre et al., 1986). On June 30 the delay trigger setting meant that these oscillations caused data records to be truncated by the range window.

ADDAAS samples the pitch and roll of the CV-990 at 1 second intervals, to a resolution of 0.01°. McIntyre et al. (1986) display typical records of pitch and roll during level flight for data collection. The standard deviation of the pitch variations is about 7′ arc, while that of the roll is 17′ arc, indicating less stability. In normal flight, the aircraft attitude is with the nose up between 1 and 3°. This attitude is determined by factors which include gross weight, the amount of fuel on board, the altitude and airspeed, and so vary throughout each flight. A typical value is 1.5° and so the sled upon which the antennas are mounted is inclined at an angle of 1.5° to the fuselage to compensate.

Platform instabilities affect the summation and tracking of return pulses. As noted in the section on tracking, aircraft motion is an important consideration, causing range noise when low PRF’s are used. Rapid terrain changes between received pulses, accentuated by inter-frame data gaps in records, cause larger errors in tracking and averaging.

Variations in pitch and roll are critical to the positioning of the illuminated footprint on the ground, and in distorting of return pulse shape. Increasing the local incidence angle and the illumination characteristics by perturbations in aircraft attitude causes changes in the apparent backscatter characteristics. As seen in the next two chapters this has important consequences for scattering and reflexion, and thus the extraction of geophysical information from recorded waveforms. Clearly altimetry data sections obtained during straight and level flight (ie. free from banking or turns) must be used to reduce errors in interpretation.

2.3.6 Geo-referencing the altimetric data

Two navigation systems were included as part of the ADDAS package, so that the flightline coordinates could be recorded on magnetic tape. These data may then be used in geo-referencing the altimetry where cloud conditions prevent the use of coincident nadir photographs.

The first system is a radio navigation system called OMEGA, which operates in the frequency range 10-14 kHz. Transmitting ground stations send long wavelength signals up to 5000 miles to be received by the aircraft; thus enabling a spatial or positional fix trigonometrically between 3 or more stations. Accuracies may better one mile, but atmospheric propagation errors are variable and unpredictable. The nearest transmitter of OMEGA navigational signals was in Norway.
The second system is the Inertial Navigation System (INS). Since this system is the more accurate of the two, these data are used almost exclusively. INS systems are based upon the principle that precise measurements of acceleration and rotation may be used to monitor the rate of change of direction of the vertical (i.e. velocity of the aircraft). The technique of integrating velocity with time then yields the absolute change in position of the aircraft (Kayton, 1969). The accuracy of the INS, however, depends upon the quality of the velocity data. Errors occur because of biases in the accelerometers, and incorrect ‘initial-conditioning’ when resetting the system before each flight. Such errors are readily propagated with time. Typical sources of errors are discussed in more detail by Kayton (1969).

Of paramount importance is the absolute accuracy of the flightline itself when locating the position of the altimeter footprint swath over the ice surfaces under consideration. In the case of the flight sections over Svalbard persistent cloud cover has precluded the use of the nadir coincident photography. This meant that to locate the flightline sufficiently well upon other forms of supporting images, an accurate corrected flightline was needed. The process undertaken was to calculate and analyse the errors, and relocate the flight track by redistribution of these errors through time. Inherently the largest component of the error is a linearly increasing one, caused by gyro drift. It may be calculated by ‘closing the loop’. This entails calculating the latitude and longitudinal errors at a known ground position; normally the airfield at the start and finish of the flight. This may be transformed into Northing and Easting distance errors, which are then distributed linearly throughout the flighttime. At each record along track an independent error must be individually calculated by changing the distance errors into latitudinal and longitudinal components for a known latitude.

2.4 Radar theory and calibration

This section presents a brief discussion of radar theory as it applies to the RAL altimeter and the measurements to be presented in later chapters. The radar altimeter fits into a broad class of devices which operate in the microwave region of the frequency spectrum, but has capabilities far more sophisticated than the original ranging and detection radars. Since it is an ‘active’ sensor (i.e. providing its own illumination) it is used to determine the backscatter coefficient, or scattering cross-section per unit area ($\sigma^o$) of the targets illuminated. The backscatter coefficient may be determined providing the detector output may be calibrated. This gives the instrument the capability to distinguish between targets with different scattering properties as well as measuring range accurately.
2.4.1 The radar equation

The radar equation relates the power transmitted to the power received for an active system. One form of the radar equation is:

\[ P_r = \frac{P_t G_t}{(4\pi R^2)^2} \sigma A_e, \]  \hspace{1cm} \{2.11\}

where \( P_r \) is the received power, \( P_t \) is the transmitted power, \( G_t \) is the gain of the transmitting antenna in the direction of the target, \( R \) is the range to the target, \( \sigma \) is the effective backscatter area of the target, and \( A_e \) is the effective receiving area of the receiving antenna.

The effective area of a receiving antenna is directly proportional to its gain \((G_r)\) multiplied by the square of the wavelength \((\lambda)\) at which the antenna is operating:

\[ A_e = \frac{G_r \lambda^2}{4\pi}. \]  \hspace{1cm} \{2.12\}

Substituting 2.12 into 2.11 yields

\[ P_r = \frac{P_t G_t G_r \lambda^2}{(4\pi)^3 R^4} \sigma, \]  \hspace{1cm} \{2.13\}

but since the RAL altimeter is a monostatic radar the value of the gain of the antenna used for transmitting and receiving is the same and 2.13 may be modified to give

\[ P_r = \frac{P_t G^2 \lambda^2}{(4\pi)^3 R^4} \sigma. \]  \hspace{1cm} \{2.14\}

Assuming that the illuminated area is composed of a large number of individual scattering components with random phase, the scattering cross-section for each individual area element on the surface \((\sigma)\) may be calculated. The average scattering cross-section per unit area, or backscatter coefficient \((\sigma^o)\) defined by Goldstein (1950) is given by

\[ \sigma^o = \frac{P_r (4\pi)^3 R^4}{P_t G^2 \lambda^2 A_{ill}}, \]  \hspace{1cm} \{2.15\}

where \( A_{ill} \) is the area illuminated on the ground.
2.4.2 The radar equation and calibration

If pointing errors and the surface height probability distribution are considered negligible, then equation 2.14 may be applied to a returned echo (Brown, 1977; and Hayne, 1980). For a delay $t$ after the first arrival of backscattered energy (i.e. in the region of the pulse), the returned power is given by

$$P_r(t) = \frac{P_t G^2 \lambda^2}{(4\pi)^3 H^4} \left(\pi H c \tau\right) \sigma^0 \exp(-\delta t),$$  \hspace{1cm} \text{(2.16)}

where $A_{ill}$ in 2.15 is related to the pulse-width $\tau$ such that the area of the surface illuminated at time $t$ is $\pi H c \tau$. Measurements made in the trailing edge of a pulse must be corrected for the shape of the antenna beam. Thus, $\delta$ is the gain pattern of the antenna, and is related to the antenna full beam width ($\theta_w$) between half power points by

$$\delta = \frac{\ln(4)}{\sin^2(\theta_w/2)} \frac{c}{H}. \hspace{1cm} \text{(2.17)}$$

Received power may be integrated up to a limiting delay, i.e.

$$\int_{t_0}^{t_n} P_r(t) \, dt = A, \hspace{1cm} \text{(2.18)}$$

where $A$ represents the area of a measured pulse shape up to a limiting delay $t_n$. Expression 2.16 may therefore be integrated to obtain

$$A = \frac{P_t G^2 \lambda^2}{(4\pi)^3 H^4} \left(\pi H c \tau\right) \sigma^0 \frac{1 - \exp(-\delta t_n)}{\delta}. \hspace{1cm} \text{(2.19)}$$

Providing the surface height probability distribution and the pulse-width are narrow in relation to the limit of integration $t_n$, then $A$ in 2.18 may be equated with the sum of power in the range bins of a recorded waveform. The mean value of power in the trailing edge of a pulse, corrected for the antenna beam pattern is thus defined as

$$\bar{P}_r = A \left[\frac{\delta}{1 - \exp(-\delta t_n)}\right]. \hspace{1cm} \text{(2.20)}$$

A value of calibrated backscatter may then be calculated using a single power value at a delay time $t$, or the integral of power in a number of range bins ($A$) in a waveform trailing edge using the following procedure. Whichever scheme is chosen, the logarithm of the power value is substituted to express values in units of dB. The following two expressions represent the form used to calculate calibrated backscatter using a single power value, and a power integral, respectively:

$$\sigma^0 = 10[\log_{10} P_r(t) - \log_{10} P_t] + 10\log_{10}\left(\frac{(4\pi)^3 R^4}{G^2 \lambda^2} \frac{\exp(\delta t)}{\pi H c \tau}\right), \hspace{1cm} \text{(2.21)}$$

$$\sigma^0 = 10[\log_{10} A - \log_{10} P_t] + 10\log_{10}\left(\frac{(4\pi)^3 R^4}{G^2 \lambda^2} \frac{1}{\pi H c \tau} \frac{\delta}{1 - \exp(-\delta t_n)}\right). \hspace{1cm} \text{(2.22)}$$
2.4.3 Operational calibration of the instrument

The backscatter measurement is derived from a direct application of the radar equations in 2.21 and 2.22 given a recorded waveform. All the parameters in these equations are known, except for accurate calibrated values of the transmitted signal power, and the received signal power. Two types of calibration are necessary for a quantitative measurement of the amplitude of the backscattered power, namely; internal and external calibration.

Internal calibration may be achieved by injecting a sample of the transmitted signal into the receiver after it has passed through a delay line of fixed length and known attenuation. This procedure is followed frequently using the inbuilt calibration mode of the RAL altimeter, to derive the ratio of transmitted power to received power, and is independent of short-term fluctuations of internal system parameters such as; cable losses, switch losses, and amplifier gains. The internal calibration loop does not include the antenna or antenna cable, so any variations in this part of the system are not taken into account by the internal calibration.

External calibration is necessary to account for drifts or variations in the internal system parameters. This procedure involved flying over a passive transponder, or 'retro-reflector' with known scattering coefficient, against which to compare the measured backscatter. The retro-reflectors are of square corner type (McIntyre, et al, 1986), and give a significant return independent of incidence angle. Nonetheless, recorded signals vary slowly in amplitude with incidence angle due to a variation in the projected area on a plane perpendicular to the incident radiation. During each overpass the aircraft's position and attitude are recorded to account for the reduction in antenna gain at the actual position of the retro-reflector in the antenna beam pattern.
2.4.3.1 Calibration for June 28, 1984

Only one internal calibration file is available near to the flight section over Nordaustlandet, Svalbard, due to faulty calibrations and data recording problems. It was made at a convenient time, directly after the final flight leg was completed over Vestfonna. Since no other internal calibration file is available from before the flight sections concerned, the assumption is made that drift in the receiver gain is negligible for the one hour operating period prior to data collection over Svalbard. Drift in receiver gain has been observed by Ulander (1985) in a two hour flight section over Greenland. The 2 dB change over this period suggests that the uncertainty in results analysed for this day is approximately ±1 dB for the 1 hour period during which data were collected. Results of the single calibration file are nonetheless given in Table 2.5 to indicate the calibrated power value and the IF attenuator setting.

Table 2.5 Results of internal calibrations relevant to data analysed from June 28 and June 30.

<table>
<thead>
<tr>
<th>Date Time</th>
<th>June 28</th>
<th>June 30</th>
<th>June 30</th>
<th>June 30</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>08.45</td>
<td>10.30</td>
<td>11.27</td>
<td>RR</td>
</tr>
<tr>
<td><strong>IF attenuator setting (dB)</strong></td>
<td>47.0</td>
<td>45.0</td>
<td>45.0</td>
<td>45.0</td>
</tr>
<tr>
<td><strong>Calibrated Power (dBm)</strong></td>
<td>-15.63</td>
<td>-13.3</td>
<td>-14.2</td>
<td>-12.8</td>
</tr>
</tbody>
</table>

An external calibration was made during the homewardbound retro-reflector crossing at Røyken on the island of Andøya. At the point of closest approach to the retro-reflector the mean signal was recorded. This and several other parameters are listed in Table 2.6. A reduction in gain is noted due to the offset of the retro-reflector from the antenna boresight axis, caused by a displacement of the aircraft flightline from a point vertically above the target. The correction is calculated at less than 0.1 dB (Wrench; personal communication, 1985)

2.4.3.2 Calibration for June 30, 1984

Two internal calibration files are available, both from the return flightleg, and shortly after the period of data collection. Results given in Table 2.5 at 10.30 and 11.27 hrs indicate a drift in receiver gain of the same order as that inferred for June 28. A difference of 0.9 dB in 1 hour indicates that the drift is not negligible over the two hour period during which sea ice data were collected. It is possible using the observed trend to interpolate the probable value of receiver gain for the period concerned, but optimally calibration files from before and after
the data section should be used. Since these are not available the accuracy of calibration is restricted to approximately ±2 dB.

External calibrations were undertaken successfully on both the outbound and inbound flights over Andøya. The latter of the two is used here for calibration since the aircraft was not off track when it crossed the retro-reflector, and so corrections in gain due to displacement from the centre of the antenna pattern are minimal (c. ±0.1 dB). The recorded parameters for this overpass are listed in Table 2.6.

Table 2.6 Results of external calibrations from Røyken retro-reflector overpasses on Andøya, for June 28 and June 30.

<table>
<thead>
<tr>
<th>Date</th>
<th>June 28 (inbound)</th>
<th>June 30 (inbound)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Range to retro-reflector (m)</td>
<td>5496.0</td>
<td>5535.0</td>
</tr>
<tr>
<td>IF Attenuator</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Setting (dB)</td>
<td>6.0</td>
<td>10.0</td>
</tr>
<tr>
<td>Mean recorded signal power to retro-reflector (dBm)</td>
<td>-14.99</td>
<td>-14.0</td>
</tr>
<tr>
<td>Gain reduction (dB)</td>
<td>&lt; 0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>Backscatter coefficient of retro-reflector (dB/m²)</td>
<td>49.9</td>
<td>52.0</td>
</tr>
</tbody>
</table>

2.4.3.3 An algorithm for routine calculation of \( \sigma^o \)

One of the aims of the following analyses is to observe the variation in \( \sigma^o \) over selected areas of Nordaustlandet, Svalbard, and Fram Strait sea ice. The scattering cross-section for a point target such as a retro-reflector, at normal incidence (ignoring corrections for the gain pattern) in a simplified form of equation 2.21, is

\[
\sigma = 10[\log_{10} P_r(t) - \log_{10} P_l] + 10\log_{10} \left( \frac{(4\pi)^3 R^4}{G^2 \lambda^2} \right). \tag{2.23}
\]

If the system response at any time during the flight is considered similar to that during a retro-reflector overflight, then the recorded power may be compared to the point target response recorded from the external calibrations. Differences in recorded power are directly attributable to changes in the backscattering properties of the surface and the retro-reflector, and variations in range to the target. The following equation is thus defined to calculate the difference in backscatter between the two targets,

\[
(\sigma_t - \sigma_{RR}) = (P_t + IF_t) - (P_{RR} + IF_{RR}) - A + B + C \tag{2.24}
\]
where: \( \sigma_t \) is the backscatter cross-section at time \( t \); \( \sigma_{RR} \) is the backscatter cross-section from the retro-reflector; \( P_t \) is the mean power at time \( t \); \( IF_t \) is the IF attenuator setting at time \( t \); \( P_{RR} \) is the mean power recorded over the retro-reflector; and \( IF_{RR} \) is the attenuator setting for the retro-reflector overflight. \( A, B, \) and \( C \) are correction factors where \( A \) is the gain reduction due to the offset of the retro-reflector, \( B \) is the height dependence

\[
B = 10\log_{10}\left( \frac{H_t^4}{H_{RR}^4} \right),
\]

and \( C \) is a correction for the drift in antenna gain.

In the case of the June 28 data, since \( A \) is less than a 0.1 dB difference it may be considered negligible and ignored. Additionally, \( C \) is incalculable as no calibration file is available at a time preceding the Nordaustlandet flight legs. The algorithm used for routine calculation of backscatter over Nordaustlandet is thus;

\[
\sigma_t = \sigma_{RR} + (P_t + IF_t) - (P_{RR} + IF_{RR}) + 10\log_{10}\left( \frac{H_t^4}{H_{RR}^4} \right). \tag{2.26}
\]

For June 30 the equation becomes slightly more complicated, as the internal calibration files are used to interpolate the receiver gain over the flight period. Since the correction for drift of the antenna gain is included, the equation becomes

\[
\sigma_t = \sigma_{RR} + (P_t + IF_t) - (P_{RR} + IF_{RR}) - A + 10\log_{10}\left( \frac{H_t^4}{H_{RR}^4} \right) +
\]

\[
+ (P_{Rcal} + IF_{Rcal}) - (P_{cal}(t) + IF_{cal}(t)) \tag{2.27}
\]

where \( P_{Rcal} \) and \( IF_{Rcal} \) are the calibrated power and the IF attenuator setting respectively for the retro-reflector overflight (given in Table 2.5). \( P_{cal}(t) \) and \( IF_{cal}(t) \) refer to the interpolated values at time \( t \) of the internal calibration signal power and attenuator setting (using the values at times 10.30 and 11.27).

The final step is to calculate the backscatter coefficient using the scattering cross-section calculated using 2.26 and 2.27. It is necessary to assume that the point target response is Gaussian as a function of time. For the RAL altimeter the full 3 dB pulse width was 7.3 ns. The slope must also be assumed negligible. This is true of most surfaces encountered on Nordaustlandet, and all of the sea ice legs on June 30.

A coefficient of backscatter expresses the total backscatter of an average point scatterer on the surface:

\[
\sigma^o_t = \frac{\sigma_t}{A_{ill}}, \tag{2.28}
\]

which is equivalent to
\[ = \sigma_t - 10\log_{10}(A_{ill}). \]  \hfill \{2.29\}

Signal power originates from a finite area for a pulse limited altimeter. Assuming a rectangular point target response, \( A_{ill} \) the area of a footprint annulus in 2.15 is given by \( \pi H c \tau \) where \( \tau \) is the full 3 dB pulse width of 7.3 ns. But the above algorithms assume a Gaussian point target response. Ulander (1985) finds the correction of \( A_{ill} \), from an area illuminated with a rectangular point target response to a Gaussian response. The new area \( A'_{ill} \) is simply

\[ A'_{ill} = 1.06(\pi H c \tau). \]  \hfill \{2.30\}

### 2.4.4 Relative and absolute calibration

Calibration enables precise measurements of the backscatter coefficient and permits accurate measurements. The difference between precision and accuracy in this case is important. Precision relates to the relative calibration of a measurement, whilst accuracy relates to the 'absolute' calibration of the same measurement. If calibration permits repeatable measurements the system has good relative calibration. If the measurements are not only repeatable but are in addition absolute then they may be considered accurate.

The monostatic radar altimeter may be used to measure the scattering or reflective properties of the surface and/or volumes within its illuminated footprint. Providing the system has been calibrated accurately internally and externally then it may yield accurate, as well as precise measurements of the backscattering cross-section, or coefficient of backscatter \( (\sigma^\circ) \). For the RAL altimeter good relative calibration is all that is necessary if one merely wishes to study time-space variations in \( \sigma^\circ \), as the same system is being used for all measurements. A bias error which may exist is the same for all times and spatial positions, and so still enables precise distinctions to be made between the characteristics of the target areas at different places at different times. Provided the relative calibration is maintained then distinctions can be made between the responses from different targets. Relative calibrations of the system must, therefore, be made regularly to account for changes in its parameters such as drift in the gain of the receiver. However the inbuilt mode enabling in-flight calibration of the instrument has been shown to be accurate during tests. Since this may be considered absolute calibration then accurate measurements of the coefficient of backscatter are possible.

Calibration files occur regularly throughout the data, enabling biases and the absolute gain of the receiver to be accounted for. The inbuilt calibration mode enables accurate internal calibration of the ratio of received to transmitted power, while external calibration of the same ratio is facilitated by flights over retro-reflectors with known scattering responses.
CHAPTER 3
REFLECTIVE PROPERTIES
OF SNOW AND ICE MEDIA

3.1 Introduction

A comprehensive outline of the electromagnetic and reflective properties of planar snow and ice surfaces is necessary before being able to interpret data records from active microwave instruments. Here a systematic treatment is made of the conditions likely to have been encountered by the RAL altimeter during flight data analysed in later chapters. Dielectric mixture formulae are used to model the electromagnetic properties governing reflexion of altimeter pulses from a variety of surface media. These enable combinations of ice, air, water, and brine to be considered. The reflexion coefficient of a smooth planar surface can be determined using Fresnel’s equations. The magnitude and phase of the reflexion depends upon, frequency, polarisation, incidence angle, and the complex dielectric constant of the surface medium (assuming the material is homogeneous). Fresnel derives two distinct equations for the cases of an electric vector either perpendicular to, or in the plane of incidence, corresponding to ‘horizontal’ or ‘vertical’ polarisations. However, the situation is simplified in the case of an altimeter, as a monostatic radar only receives energy by plane reflexions from surface facets oriented perpendicular to incident radiation. At normal incidence the two polarisations become identical and so the following discussion excludes incidence angles other than 90°.

Consideration must be given to the electromagnetic properties of snow and ice media in order that the variability of reflexion from the variety of surfaces encountered by a radar altimeter in polar regions is accountable. The dielectric properties of snow and ice are known to cover a wide range of values (Von Hippel, 1945; Johari, 1981). They depend primarily upon the density, temperature, free water content, and impurity content, but important secondary effects result from sequences of rain, and the diurnal variations which may cause freeze and thaw cycles. There are also other important practical considerations such as the effects of surface roughness, but this initial discussion is a theoretical study concentrating upon planar surfaces only, and such complications are treated in the following chapter. The basic objective is to present results of calculations made using a variety of models, and to answer the basic question; for sets of snow and ice conditions both on land and sea, how is the reflexion coefficient determined by the dielectric properties of the medium?

Importantly, there is a lack of experimental results at 13.8 GHz, for polar snow and ice
surfaces. Cases where corroborative evidence is available are reported, but in the case of saline sea ice, experimental evidence is limited to 10 GHz. Since the majority of sea ice dielectric mixture models are empirically derived, results discussed are restricted to that frequency. Until further work is conducted for the full range of GHz frequencies, these results may be used as a first order estimate of the relevant dielectric properties and reflection coefficient of saline ice.

3.2 Theoretical considerations

A schematic diagram of electromagnetic interactions is illustrated in Figure 3.1 for the simplified case of a single dielectric interface and a non-normal angle of incidence. For the moment, what happens to the transmitted energy is not investigated, and only interaction at the air/ice or air/snow interface is considered. In the following chapter the theory used to describe interactions at the air/medium may be applied to internal boundary reflection. Thus what happens to the penetrating energy is not ignored; this treatment simply discusses reflection at the upper surface of a single layer medium.

Figure 3.1 Reflection at a dielectric interface between medium 1 (air) and medium 2 (specified medium). $I_o$ is the incident energy, $I_r$ is the reflected energy, and $I_t$ is the transmitted energy. The density of each medium is $\rho$, $\epsilon'$ and $\epsilon''$ are the real and imaginary parts of the dielectric constant, and $\mu_0$ is the permeability of the medium.

Smith (1971) outlines the theory behind Fresnel reflection at an interface between two specified media. He demonstrates that the intrinsic impedance $\eta$ of the media govern reflection at a distinct boundary or dielectric interface. The amplitude reflection coefficient $\rho_{12}$ for an
electromagnetic wave impinging upon the boundary between medium 1 (air) and medium 2 is simply

$$\rho_{12} = \frac{\eta_2 - \eta_1}{\eta_2 + \eta_1}, \quad \{3.1\}$$

where $\rho_{12}$ is the amplitude reflexion coefficient, and $\eta_1$ and $\eta_2$ are the respective impedences. Additionally the transmission coefficient $\tau_{12}$ may be calculated, and is

$$\tau_{12} = \frac{2\eta_2}{\eta_2 + \eta_1}. \quad \{3.2\}$$

These two coefficients may be expressed as a power reflexion coefficient $R_{12}$ and a power transmission coefficient $T_{12}$ indicating the proportions of incident power either reflected from or transmitted across the interface;

$$R_{12} = \left(\frac{\eta_2 - \eta_1}{\eta_2 + \eta_1}\right)^2 \quad \{3.3\}$$

$$T_{12} = \left(\frac{2\eta_2}{\eta_2 + \eta_1}\right)^2. \quad \{3.4\}$$

Of additional note is the fact that the sum of power reflexion and transmission coefficients is unity, ie. energy is conserved at the interface.

The intrinsic impedance $\eta$ is a complex quantity defined by the following formula;

$$\eta = \sqrt{\frac{\mu_0}{\varepsilon_0} \left(\frac{1}{\varepsilon^*}\right)} \quad \{3.5\}$$

where $\mu_0$ is the permeability of free space, $\varepsilon_0$ is the permittivity of free space, $\varepsilon^*$ is the complex dielectric constant of the medium, and the permeability of the medium $\mu$ is assumed to be 1. The complex dielectric constant may be expressed in terms of the permittivity $\varepsilon$ and conductivity $\sigma$ of a medium by;

$$\varepsilon^* = \left(\frac{\varepsilon}{\varepsilon_0}\right) - j\frac{\omega \sigma}{\varepsilon_0} \quad \{3.6\}$$

where $\omega$ is the angular frequency (ie. $2\pi f$ rad s$^{-1}$), or more simply in terms of the relative permittivity $\varepsilon'$ and dielectric loss $\varepsilon''$;

$$\varepsilon^* = \varepsilon' - j\varepsilon'', \quad \{3.7\}$$

where permittivity and dielectric loss are real and imaginary parts of the complex dielectric constant.

Equation 3.5 may be expressed in terms of the loss tangent $\tan \delta$ (where $\tan \delta = \varepsilon''/\varepsilon'$), which characterises the absorptive properties of the medium;

$$\eta = \left[\frac{\mu_0}{\varepsilon_0} \left(\frac{1}{\varepsilon'(1 - \tan \delta)}\right)\right]^{\frac{1}{2}}. \quad \{3.8\}$$
From equation 3.6 it is evident that $\varepsilon^*$ is frequency dependent, owing to the frequency dependence of the dielectric loss ($\varepsilon''$) in the microwave range. For media which have negligible losses the real part of the complex dielectric constant dominates. The converse situation occurs with high loss or 'lossy' dielectrics, where the imaginary part of the complex dielectric constant becomes significant. As a rule of thumb low loss dielectrics are those where $\tan \delta \ll 1$.

Variations in $\varepsilon^*$ are largest in the interval of the relaxation frequency (ie. the frequency where dielectric losses are greatest), where absorption is caused by the relaxation spectrum of the medium. For pure polycrystalline ice, this frequency is in the kHz region or infrared band, while that of water at 0°C is at 8.5 GHz (Rott et al., 1985). The relaxation frequencies of heterogeneous polar materials such as wet snow, for instance, tend to be higher than for its individual components ice and water (De Loor, 1964; Ambach and Denoth, 1972). This complicates the description of the dielectric properties of mixtures.

Those types of surface likely to be encountered by the RAL altimeter are discussed individually in the following sections. Where heterogeneous substances are treated, relevant dielectric mixture formulas are applied. This enables calculation of the likely electromagnetic properties of a variety of media over typical ranges of temperature, density, and salinity, and thus their reflexion coefficients.

### 3.3 Reflexion from freshwater ice and dry snow

#### 3.3.1 Ice as a low loss dielectric

The relative permittivity $\varepsilon'$ of ice may be considered constant throughout the GHz range (Lamb, 1946; Von Hippel, 1945), and in fact the whole microwave range (Johari,1981). Various workers have conducted experiments and have derived a value of $3.15 \pm 0.02$ (Cumming, 1952; Blue, 1980; and Stiles and Ulaby, 1982) for ice without impurities. According to Paren (1970) ice has corresponding conductivity of around $3.0 \times 10^{-5} \Omega^{-1}m^{-1}$ at $-10^\circ C$ which decreases with temperature. Experimental measurements of the dielectric loss ($\varepsilon''$) of ice have also been made between 1 and 30 GHz (Johari, 1981). It remains very low throughout the microwave frequency range with values reaching a minimum of less than $10^{-3}$ at 10 GHz, for ice at $-5^\circ C$. Ice may therefore be considered a low loss dielectric, having negligible losses at 13.8 GHz, the frequency of the RAL altimeter. Substituting $\varepsilon'$ in place of $\varepsilon^*$ in equation 3.5 gives a good approximation of the intrinsic impedance of any low loss medium. A value of 3.15 for $\varepsilon'$ gives a power reflexion coefficient $R$ of 0.078 for a pure ice surface (from equation 3.3). This shows that over 90% of incident energy is transmitted across the dielectric interface. The amount of power reflected expressed in decibels is approximately -11.0 dB. This is, however,
a laboratory estimate of the reflexion properties of an ice surface. In reality impurities cause higher dielectric losses and so this value of $R$ may be regarded as the maximum reflexion coefficient which freshwater ice surfaces may attain.

### 3.4 Physical properties of snow

Snow is an aggregate of air and of different shaped grains of ice. This ice/air ratio is variable and governs the density of the snow medium. If all air is excluded the medium may be regarded as pure ice. This limit is generally accepted as occurring at a density of approximately 918 kg m$^{-3}$ in pure glacier ice.

Snow grains may be single ice crystals or several cohering particles of ice, but as such are recognised as the mechanically separate unit of the snowpack. The density of newly fallen snow is generally about 100 – 300 kg m$^{-3}$ depending upon the ice crystal size and shape, wind velocity, and terrain features (Mellor, 1964). Initially snow particle morphology depends upon the temperature and humidity of the atmosphere through which the snow falls. Newly deposited snow is a loosely packed aggregate of intricate dendritic and platelike ice crystals. If deposition occurs in strong winds then crystals may be broken and shattered into small pieces which consequently become more densely packed.

Metamorphism of the snow cover begins immediately following deposition. The natural tendency of thermodynamic processes is to reduce surface free energy through phase changes, in this case minimising the ratio of surface area to volume of ice particles. The ideal shape to achieve this minimum is a sphere, and so in time even the most intricate snow crystals tend to become rounded particles of ice. Transfer of water molecules from one part of a crystal to another through the vapour phase (ie. sublimation and redistribution as a solid) is the dominant mechanism. Such destructive metamorphism, as it is known, is most rapid at temperatures close to freezing point and diminishes in intensity as temperature falls. Below about $-40^\circ$C it virtually comes to a halt. Destructive metamorphism first transforms newly fallen snow into a more stable settled layer of rounded ice particles of density 200 – 300 kg m$^{-3}$, with typical grain diameters between 0.5 – 1.0 mm. Sublimation continues to occur and larger grains form at the expense of smaller particles at a decreasing rate. Destructive processes only occur more rapidly at higher ambient temperatures. Furthermore, ‘melt metamorphism’ can accelerate rounding of snow grains: melt-freeze cycles aid the redistribution of mass in the form of free water. Grain growth may take place up to about 3 mm diameter, and the snow becomes more densely packed (Seligman, 1937). Destructive metamorphism is complete when snow particles have been reduced to rounded grains (ideally spheres) of ice. Such snow normally has a density of 500 – 600 kg m$^{-3}$. The bulk density of randomly nested ice spheres for comparison is about 580 kg m$^{-3}$. 

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Further density increases may only result from a process known as 'firnification'. The two main mechanisms are compaction of the snow matrix under pressure, and refreezing of meltwater which is trapped in interstices by capillary action. However, the former process need not be discussed in this context since it may only occur at depth in a snowpack. The latter is typical of spring and early summer as the snow surface layers experience fluctuations in temperature and thus frequent melting and freezing. This can cause surface layers of densities between 600 kg m\(^{-3}\) and solid ice at 918 kg m\(^{-3}\).

### 3.4.1 Dry snow mixture dielectrics

Dry snow is a heterogeneous substance and can be considered a mixture of air and ice. There exists a continuum of states between on the one hand low density snow with a large proportion of air, and solid polycrystalline ice with no air bubbles at all. This has enabled simple models to be formulated, which can be used to calculate the complex dielectric constant from the relative proportions of ice and air, or more simply from its density. The dielectric properties of ice are well understood, enabling dielectric mixture formulae to be employed to give the bulk dielectric constant of a snow volume.

Several mixing formulae have been derived to describe its dielectric or conductive properties. Tiuri et al. (1984) derive a simple empirical relationship which enables the calculation of the relative dielectric constant of dry snow, given its density. Paren (1970) considers dry snow as a mixture of ice and air in proportions defined by the ratio of snow density (\(\rho_s\)) to ice (\(\rho_i\)); where \(\rho' = \rho_s / \rho_i\). The following empirical relationships are used to calculate the dielectric constant \(\epsilon_s\) and conductivity \(\sigma_s\) of the snow:

\[
\begin{align*}
\epsilon_s^{1/3} - 1 &= \rho' \left(\epsilon_i^{1/2} - 1\right) \\
\sigma_s &= \sigma_i \rho' (0.68 + 0.32 \rho')
\end{align*}
\]

Despite being able to calculate the conductivity of the snow medium, its value remains negligible for dry ice/air mixtures. Dielectric losses as a result are negligible and so only the real part of the complex dielectric constant calculated from equation 3.9 need be substituted in equation 3.5 to calculate the intrinsic impedance of dry snow. Figure 3.2 demonstrates how dry snow density controls the dielectric constant and reflexion coefficient at an air/snow interface, using Paren's equations. The reflexion coefficient is expressed in dB where \(R_{dB} = 10 \log_{10}(R_{12})\) from equation 3.3. Experimental values by Cumming (1952) are included to show that the results from this dielectric mixture model correspond closely with reality. In general, \(\epsilon'\) increases almost proportionately with density between 1.25 and 3.15, over the normal range of snow densities encountered. In contrast, the reflexion coefficient increases
Figure 3.2  Variation in the mixture formula results for the relative permittivity \( \varepsilon' \) of a dry snow medium, and the corresponding reflexion coefficient \( R \). Experimentally derived values for \( \varepsilon' \) are plotted to indicate the accuracy of the dielectric mixture formula of Paren (1970).

Exponentially, rising from -22.5 dB at 200 kg m\(^{-3}\) to -11.0 dB at the density limit of 918 kg m\(^{-3}\) for pure ice.

The main drawback with equations 3.9 and 3.10 is that they do not take into account the structure of the snowpack. Theoretical models of mixtures of dielectrics generally consider a distribution of separate ice particles, of idealised shape, within a dielectric medium of air. One example is Polder and Van Santen’s (1946) formula for ellipsoidal particles of a solid dielectric. However, a more simple approach is adopted here, as this type of theoretical approach is not strictly applicable in the context of complex ice grain shapes. Wiener (1910) shows empirically that a single parameter \( u \), called the ‘formzahl’ or form number is sufficient to describe how one dielectric is dispersed within another. Figure 3.3 is a schematic diagram illustrating how Evans (1965) relates snow structure to the formzahl \( u \), when an electric field normal to the surface is considered. He draws attention to the fact that the dielectric constant is not only a function of the relative proportions of the two media but also of the aggregation and orientation of ice particles. It is shown experimentally that if particles of high permittivity are elongated in the direction of the electric field then the combined permittivity is higher than if they lie across the field, and so the mixture is anisotropic. Wiener (1910) derived the following relationship
to find the dielectric constant of a mixture \( (\varepsilon^*_m) \):

\[
\left( \frac{\varepsilon^*_m - 1}{\varepsilon^*_m + u} \right) = \rho' \left( \frac{\varepsilon^*_1 - 1}{\varepsilon^*_1 + u} \right) + (1 + \rho') \left( \frac{\varepsilon^*_2 - 1}{\varepsilon^*_2 + u} \right)
\]  \hspace{1cm} \{3.11\}

where \( \varepsilon^*_1 \) and \( \varepsilon^*_2 \) are the complex dielectric constants of the two components. \( \rho' \) is the proportion of the total volume occupied by medium 1 (if medium 1 is ice then \( \rho' = \rho_m / 0.92 \), where \( \rho_m \) is the density of the snow), and \( u \) is the form number. In the case of dry snow, as described previously, the losses are assumed negligible and so only the permittivity \( \varepsilon' \) of ice need be substituted for \( \varepsilon^*_1 \) in equation 3.11. Also, medium 2 is air: if a value of 1 is used for \( \varepsilon^*_{\text{air}} \), then the right hand side of the equation is zero and 3.11 becomes

\[
\varepsilon^*_m = \left( \frac{\varepsilon'_1 + u + \left( \rho' \varepsilon'_1 u - \rho'u \right)}{\varepsilon'_1 + u - \rho' \varepsilon'_1 + \rho'} \right).
\]  \hspace{1cm} \{3.12\}

**Figure 3.3** Illustration of the relationship between the Formzahl \( u \) and the dielectric mixture structure, for a dry snow mixture of ice and air (after Evans, 1965).

**Figure 3.4** Relative permittivity of snow versus density. The lower curves are computed using Wiener's formula with Formzahls of \( u = 0, 2, 10, 25, \infty \). The upper curves are the corresponding reflexion coefficient for a dry snow surface layer of varying dielectric structure.
In Figure 3.4 the results of calculating the relative permittivity of the mixture for different formzahls, shows how snow structure controls the reflexion coefficient. Cumming's (1952) experimental values for the permittivity of dry snow correspond with form numbers between 2 and 4. It is suggested then that in practice formzahls between 2 and $\infty$ are expected depending upon the structure of the snow. Curves for these two values in turn set limits for expected values of the dielectric constant and reflexion coefficient for snow of known specific gravity. These observations show that a maximum variation in the reflexion coefficient of around 5 dB can occur at a density of 200 kg m$^{-3}$ for different snow structures.

The loss factor $\tan \delta$ of dry snow can be calculated. The imaginary part $\varepsilon''$ of the complex dielectric constant of dry snow is

$$\varepsilon'' = \frac{1}{\rho'} \left( \frac{\varepsilon'_m - 1}{\varepsilon'_i} \right)^2.$$  

Subscript $i$ refers to medium 1 (ice) and subscript $s$ to the mixture (snow). The equation does not directly relate to the formzahl, but the permittivity of snow ($\varepsilon'_m$) may be computed for different values of $u$ with equation 3.12, or using the experimental values of Cumming (1952). Calculations made by Cumming show that the maximum value for the loss tangent of snow occurs at 0°C and a density of 918 kg m$^{-3}$ (ie. pure ice), and is $26 \times 10^{-4}$. Von Hippel (1945) calculates a similar maximum of $7 \times 10^{-4}$ at $-12$°C for pure ice. Values from calculations using equations 3.12 and 3.13 indicate that dry snow has a loss factor ($\tan \delta$) of less than half this value for densities below 500 kg m$^{-3}$, when $u = 2$. Since at a formzahl of 2, $\varepsilon'_s$ is reduced almost in proportion to snow density, the loss factor as a rough guide also falls in direct proportion to density. Evidently the density and temperature cause the greatest variation in losses, and the reflexion coefficient responds accordingly. The main complication is when impurities such as water are introduced into the snow matrix, as small proportions of liquid can have widely varying effects on the loss characteristics, and thus the reflective properties of the surface. The effects of introducing liquid water must therefore be investigated.

### 3.5 Reflexion from wet snow surfaces

At the time of year that the RAL altimeter observations were made (June/July in the Northern hemisphere) the likelihood is that snow melt may have occurred in places. The reflective properties of wetted snow at normal incidence are markedly different from dry snow surfaces, due to the significant influence of free water upon the mixture dielectrics. Variations in the reflexion coefficient are therefore investigated for differing snow water contents, densities, and grain sizes.
3.5.1 Wet snow mixture dielectrics

For wet snow the permittivity and dielectric loss are frequency dependent, showing a strong increase with liquid water content in the GHz range. Grain size, porosity, and structure to some degree also influence the dielectric properties of the medium (Sweeny and Colbeck, 1974). Many measurements have been made at MHz frequencies (Evans, 1965; Smith and Evans, 1972), but a high frequency extrapolation above 10 MHz carried out by Denoth et al. (1984) showed that the dielectric constant is only frequency independent as far as 1 GHz. Beyond this point measurements have been made by Sweeny and Colbeck (1974), Cumming (1952), and Tobarias et al. (1978), at fixed frequencies of 6, 9, and 9.4 GHz. Linlor et al. (1980) conducted experiments over the interval 4 - 12 GHz for artificially moistened snow, and investigations have been made by Hallikainen et al. (1982) of naturally moist snow in the 4 - 18 GHz range.

The presence of impurities or water in small quantities has a large effect on the complex permittivity of an ice/air mixture (Evans, 1965). This effect is of paramount importance at centimetre wavelengths (13.8 GHz ≈ 2.2 cm) when liquid water of high relative permittivity is introduced. In fact the effective relaxation frequency \( f_0 \) of undisturbed wet snow (i.e. the frequency where \( \epsilon'' \) has its maximum value) was found by Hallikainen et al. (1982) and Mätzler et al. (1984) consistently at 10 GHz ± 1 GHz. Loeb et al. (1971) have observed a relaxation frequency of 11 GHz for pure water, and so it is concluded that the dielectric properties of wet snow mixtures depend largely on the relaxation spectrum of water.

Linlor et al. (1980) derive a number of empirical relationships for the calculation of \( \epsilon' \) and \( \epsilon'' \) from their experimental work. The following formula is used to calculate the relative permittivity \( \epsilon'_{ws} \) of wet snow;

\[
\epsilon'_{ws} = 1 + 2\rho_s + yV_w^{3/2},
\]

where

\[
y = 5.87 \times 10^{-2} - 3.1 \times 10^{-4}(f - 4)^2,
\]

where \( f \) is the frequency in GHz; and \( \rho_s \) is the density of the snow. The loss factor of wet snow (\( \tan\delta \)) is calculated using the following formula;

\[
\tan\delta = \left( \frac{1.0994}{f\sqrt{\epsilon'_{ws}}} \right) x.
\]

The factor \( x \) may be calculated given the snow grain diameter \( A \), the frequency \( f \), and the volume fraction of water \( V_w \) by;

\[
x = V_w (0.045(f - 4) + 0.066(1 + A)).
\]
Equation 3.16 may be reduced to give the dielectric loss $\varepsilon''_{ws}$ of wet snow,

$$\varepsilon''_{ws} = \left( \frac{1.0994 \sqrt{\varepsilon'_w}}{f} \right). \tag{3.18}$$

**Figure 3.5**  (a) shows curves for $\varepsilon'$ and $\varepsilon''$ at three densities of snow and varying water content. Snow grain diameter is 1 mm. (b) shows corresponding reflection coefficients at each density.

Figures 3.5 to 3.7 clearly indicate what effects the variation of crystal size, density, and water content have upon the dielectrics of a wet snow medium. In Figure 3.5a, for a typical snow grain diameter of 1 mm, increases in both the permittivity and dielectric loss occur as the water content of the snow increases from 0 to 15%. Over this range of water content $\varepsilon'_w$ increases steadily by approximately 1.9 at all densities, reaching 4.1 at 15% water content and 750 kg m$^{-3}$ density. In contrast, $\varepsilon''_{ws}$ increases gradually over this range of water contents, rising to a value of 1.5 for a snow density of 750 kg m$^{-3}$. The effect of density upon the dielectric loss only becomes significant when $W_v$ is large. The difference in $\varepsilon''_{ws}$ between snow of density 500 and 750 kg m$^{-3}$, at $V_w = 15\%$ for instance, is 0.1. Figure 3.5b illustrates the effect of the dielectric properties outlined above, upon the reflection coefficient. $R$ is expressed in dB, and all the curves apply at a fixed snow grain diameter of 1 mm. As expected, the effects of small water contents upon the reflective properties are magnified for lower density snow. At a density of 200 kg m$^{-3}$ an increase of 5 dB occurs for an increase in $W_v$ from 0 to 5%. As density increases this effect becomes less pronounced. For a similar increase in water content at a density of 500 kg m$^{-3}$, $R$ only increases by 1 dB.

Figures 3.6a and 3.6b show densities of 200 and 500 kg m$^{-3}$. The effects of varying snow
Figure 3.6  (a) Model results for $\epsilon'$ and $\epsilon''$ for wet snow of varying snow grain diameter and water content. The dry snow density is $200 \text{ kg m}^{-3}$. (b) is a similar plot for an increased density of $500 \text{ kg m}^{-3}$.

The most important consequences of the observed variations, however, are that the reflexion coefficient of wet snow increases extremely rapidly with small increases in water content. The effects are most pronounced in low density snowpacks between 200 and $500 \text{ kg m}^{-3}$, and with water contents below 10% by volume. Beyond 15% by volume the reflexion coefficient approaches its maximum value. The effects of grain size upon dielectric losses are small and the resulting change in the reflexion coefficient is negligible, and of second order importance only.
Figure 3.7 Calculations of the reflexion coefficient of wet snow surfaces of varying water content and snow density. At densities a, b, and c, there are three curves which represent the calculated variation with snow grain diameter. The effects of different grain sizes manifested in Figures 3.6a and b for $\epsilon'$ and $\epsilon''$ have negligible effect upon $R$.

In snow, water saturation generally varies from 0 to around 14 % of pore volume. But, this value corresponds with the upper limit of the 'pendular regime' (Smith, 1933) which represents the dominant regime in natural snowpacks. This 14 % of pore volume corresponds to 8.4 % liquid content (by volume) in a 60 % porosity snow, and 5.6 % liquid water content in a 40 % porosity snow. Beyond this point the 'funicular regime' includes saturations up to the point at which all the pores are filled with water. However, 'pore saturations' of 14 % of pore volume only occur when drainage is impeded by an obstacle. In fact, under normal circumstances, drainage will begin at 7 % saturation of pore volume, and so one expects to encounter snowpacks with a pendular regime in the majority of situations.

The 14 % saturation limit is calculated for different densities displayed in subsequent graphs. Firstly for a dry snowpack of density 200 kg m$^{-3}$ the upper limit to the pendular regime is 11.2 % water content by volume. Secondly, for a density of 500 kg m$^{-3}$ the limit occurs at 7 % water content. Finally, for an upper density of 750 kg m$^{-3}$ the funicular regime begins at only 3.5 % water by volume. These limits may be identified for a particular snowpack to aid delineation of a range of values occurring under natural circumstances.

A problem with applying Linlor's work is that the relaxation frequency $f_o$ was found higher then 12 GHz. This indicates most importantly how the manual mixing of snow and water disturbs natural conditions. Tiuri et al. (1984) provide an alternative mixing formula for undisturbed wet snow. They utilise Taylor's (1965) dielectric mixture formula to consider
a combination of air and water between ice particles. Water inclusions in the pore spaces are considered to be halfway between randomly oriented needle-like shapes and spheroids, since this is closer approximation to what may be expected in reality. Their empirical formula derived to find the permittivity of dry snow ($\varepsilon'_{\text{dry}}$) is;

$$\varepsilon'_{\text{dry}} = 1 + 1.7\rho_{\text{dry}} - 0.7\rho_{\text{dry}}^2 \tag{3.19}$$

where $\rho_{\text{dry}}$ is the density of the snow when dry. The increase in the real part of the complex dielectric constant of snow ($\Delta\varepsilon'_{\text{wet}}$) caused by liquid water is;

$$\Delta\varepsilon'_{\text{wet}} = \varepsilon'_{\text{wet}} - \varepsilon'_{\text{dry}} \tag{3.20}$$

Tiuri et al. derive the following relationships which are applicable in the GHz frequency range;

$$\Delta\varepsilon'_{\text{wet}} = \varepsilon'_w(0.10W_v + 0.80W_v^2), \tag{3.21}$$
$$\varepsilon''_{\text{wet}} = \varepsilon''_w(0.10W_v + 0.80W_v^2) \tag{3.22}$$

where $\varepsilon'_w$ and $\varepsilon''_w$ are the real and imaginary parts of the complex dielectric constant of water, $W_v$ is the volume fraction of water, and $\varepsilon''_{\text{wet}}$ is the dielectric loss of the wet snow. $\varepsilon'_w$ and $\varepsilon''_w$ may be calculated using the following equations, which are derived from a simple Debye equation for water;

$$\varepsilon'_w = 4.9 + \left(\frac{82.8}{1+(f/f_o)^2}\right), \tag{3.23}$$
$$\varepsilon''_w = \frac{82.8(f/f_o)}{1+(f/f_o)^2} \tag{3.24}$$

where the relaxation frequency ($f_o$) of water is 8.84 GHz.

Figure 3.8a shows the results of calculations of $\varepsilon'_{\text{wet}}$ and $\varepsilon''_{\text{wet}}$ using equations 3.19 and 3.22 and by varying the water content of the snow. Most noticeable is the shallower gradient in the increase of the relative permittivity, in comparison with the results of Linlor et al.. The dielectric loss calculations show equivalent values although the equations of Tiuri et al. do not take into account variations in snow grain diameter. Figure 3.8a also illustrates the bounds of the funicular and pendular regimes. These limits may be identified for a particular snowpack to help delineate the range of dielectric properties occurring in normal circumstances under wet snow conditions. In Figure 3.8b the reflection coefficients are calculated for the three corresponding densities, and the pendular regime indicated. For the range of snow densities 200 - 750 kg m$^{-3}$, the limits to the reflection coefficient are established as -22 and -10 dB.

The main difference between the respective models of Linlor et al. and Tiuri et al. is that the former accounts for the structure of the snowpack to some degree by altering grain
Figure 3.8  (a) $\varepsilon'$ and $\varepsilon''$ for wet snow of varying density and volume fraction of water. The shaded area represents the range of water contents at each density in the 'pendular regime'. (b) shows the reflexion coefficients for the corresponding densities and water volume fractions.

diameter. Though this does not affect the real part of the dielectric constant it does increase dielectric losses the larger grain diameters become. Despite these variations the ultimate effect upon the reflexion coefficient is of minor importance. In making comparisons between these results and Tiuri et al’s results a reasonably close correspondence is found between the imaginary parts of the dielectric constant when the grain diameter (A) in the former is set to 1 mm. The most important factor is that the dielectric constant of wet snow obeys a Debye relaxation spectrum with a relaxation frequency of 10 GHz (Mätzler et al., 1984) at 0°C. Although this is consistent with the fact that the most important component of the dielectric mixture, water, has a relaxation frequency of 9 GHz at 0°C, it is not in agreement
with the findings of Linlor et al., who observe a relaxation frequency > 12 GHz for their wet snow samples. Importantly, a frequency of 13.81 GHz is above the true relaxation frequency of wet snow, and so the results of Linlor et al. give an artificially high value of permittivity. The results from these experiments, therefore, do not show full agreement (Tiuri et al., 1984) but the main differences between the two may be accounted for by the different techniques in preparation of the snow samples.

Results in this section show that wet snow can have a coefficient of reflexion which is of the order of 10 or even 15 dB higher than that of dry snow at low densities. This situation holds true for only small free water contents by volume (less than 11% at a density of 200 kg m\(^{-3}\)), when water is mainly suspended between the necks of snow particles under capillary action. The higher the density of the snowpack, however, the smaller the difference between the wet and dry situation. Comparison of the results of Linlor et al. (1980), and Tiuri et al. (1984) are also valuable in showing that disturbing the snow layer to introduce the water causes increases of the order of 1 dB in the value of \(R\). Inhomogeneity in snowpacks, in terms of crystal shape, size, and orientation, is therefore considered important in the case of wet surfaces, and may be expected to increase the relaxation frequency of the mixture and the reflexion coefficient \(R\) at the surface.

### 3.5.2 Saturated snow

Under abnormal conditions, the snowpack may become soaked and completely saturated (ie. the funicular regime). As explained, drainage must be impeded in some way, possibly by ice lenses or seasonal layers. Paren (1970) derives the following formula which may be used to find the relative dielectric constant of soaked snow (\(\varepsilon'_{\text{wet}}\));

\[
\varepsilon'_{\text{wet}} = \varepsilon'_{\text{dry}} + \frac{1}{3} \varepsilon'_{w}(1 - \rho').
\]  

(3.25)

All available air spaces are assumed to be filled with water and lie in randomly-oriented but interconnecting veins. Referring back to equation 3.9 the relative permittivity (\(\varepsilon'_{\text{dry}}\)) and \(\rho'\) may be found, and the relative dielectric constant of water (\(\varepsilon'_{w}\)) calculated from equation 3.23. The results of calculating \(\varepsilon'_{\text{wet}}\) and \(R\) are plotted in Figure 3.9, illustrating clearly how the relative permittivity decreases with density. The reason for this dependence is that the pore space decreases with density, so reducing the amounts of free water. The coefficient \(R\) tends to a maximum of 4 dB for a fully saturated snow at 200 kg m\(^{-3}\), but falls to a minimum of -11.0 dB at the limit 918 kg m\(^{-3}\) in which all available air spaces are imagined filled with pure water.
Figure 3.9 Relative permittivity $\varepsilon'$ and reflexion coefficient $R$ calculated for saturated snow. All available pore spaces are assumed to be filled with water.

3.6 Reflexion from sea ice

During MIZEX '84 several groups working from different research vessels made measurements of the physical and dielectric properties of the surface of ice floes. Burns et al. (1986) report observations of floe surface snow and sea ice dielectric properties at frequencies of 0.1, 1.0, and 10.0 GHz. The results of salinity, temperature, and dielectric constant measurements may be used along with ice property measurements by Tucker et al. (1987) to investigate the electromagnetic response of a saline ice surface. Although in the majority of cases sea ice was observed to have a surface layer of snow the ice dielectric properties are also important especially when penetration occurs through the snow. Under certain circumstances bare ice surfaces were observed, especially with first year ice. First year floes on average had only a thin layer of snow and so floes which had undergone surface melting sometimes displayed a patchy snowcover or bare ice surface.
3.6.1 Sea ice physical properties

Saline ice may be classified into two important structural categories; namely frazil and congelation ice. Frazil often forms from an agglomeration of ice particles which have grown in supercooled turbulent water, while in contrast congelation ice forms in calm water or under an existing ice cover such as frazil (see Appendix A). This classification scheme, however, is in many respects too simplistic because the presence of salts makes the problem far more complex in remote sensing terms. Salts cause sea ice to be multiphase in nature, and influence the dielectrics to a large degree. A better classificatory scheme is based on sea ice age since age determines the amounts of impurity (ie. salts) in the ice, and thus the dielectric and reflexion properties of the sea ice medium. The two most important types of sea ice are young or first year ice, which has not been subjected to summer melt, and multiyear ice, which has survived throughout one or more summer seasons (see Appendix A). First year ice has high salinity and comprises an uppermost layer of polycrystalline frazil ice with randomly oriented grains, followed by a layer of columnar crystal orientation beneath it. The shapes of brine pockets in frazil approximate to spheroids, and are elongated parallel to the growth direction in columnar ice. Salinities in first year ice in general can have concentrations between 4 – 20 °/oo (parts per thousand by mass). In contrast, multiyear ice undergoes substantial melting, refreezing, and brine drainage during the summer season, and has lower salinities, often less than 1 ppt. Additional melting causes the ice to take on a gently rolling topography, consisting of hummocks and meltponds.

Observations made by Tucker et al. (1987) during MIZEX '84 indicate that the main distinction between multiyear and first year ice floes was on the basis of salinity in the upper layers. Figure 3.10 shows the mean salinity profiles of these two ice types up to a depth of 2 m for the month of June. These indicate that the mean salinity in the upper 2 m of multiyear ice is approximately 1.5 °/oo and that of first year ice is 4°/oo. Importantly the profile shows that cycles of elevated air temperature and increased solar radiation have already caused recrystallisation and reductions in brine volumes in the upper layers of the first year ice by late June (reduced by 3 °/oo in upper 30 cm). In several unusual cases extremely low salinities (< 1 °/oo) were observed in the upper few centimetres of first year floes (Tucker et al., 1987). One suggested explanation is that such low salinity layers are composed of refrozen snow or frozen rainwater, as they are observed to be composed of crystals typical of frozen freshwater. In contrast, unnaturally high salinities were also observed in extreme cases on the surfaces of multiyear ice floes. Values as high as 3.5 °/oo or more result from wave washing and flooding by seawater.

Sea ice is thus a heterogeneous mixture of brine pockets within an ice matrix. The brine pockets contain salt and liquid water. To understand its dielectric properties the dielectric
properties of the main components, ice and brine, must be investigated.

Figure 3.10  Mean salinity profiles for Greenland Sea marginal ice zone ice from observations during MIZEX '84 (Tucker et al., 1987). FY is Firstyear Ice and MY is Multiyear Ice.

3.6.2  Brine volume as a function of temperature, density and salinity

Frankenstein and Garner (1967), and Poe et al. (1972) have derived relationships between brine volume and salinity and temperature of sea ice assuming a fixed ice density of 926 kg m\(^{-3}\). The relative brine volume \(V_b\) has been investigated by Assur (1960) and is the basis for the empirical equations of Frankenstein and Garner (1967) discussed below. Typical values of density of first year ice are observed to vary between 800 kg m\(^{-3}\) and 960 kg m\(^{-3}\) for the normal range of salinities found. In a single ice floe the density is relatively uniform, varying only slightly with temperature, salinity, and porosity. So, for the accuracy required in most calculations the assumption of constant density should not be too critical.
Poe et al. (1972) give the following relations for the brine salinity $S_b$ in parts per thousand, as computed from original data by Assur on the variation in $S_b$ with temperature $T$:

\[-8.20 ^\circ C \leq T \leq -2.0 ^\circ C \quad S_b = 1.725 - 18.756T - 0.3964T^2,\]
\[-22.9 ^\circ C \leq T \leq -8.2 ^\circ C \quad S_b = 57.041 - 9.929T - 0.16204T^2 - 0.002396T^3,\]
\[-36.8 ^\circ C \leq T \leq -22.9 ^\circ C \quad S_b = 242.94 + 1.5299T + 0.4291T^2,\]
\[-43.2 ^\circ C \leq T \leq -36.8 ^\circ C \quad S_b = 508.18 + 14.535T + 0.2018T^2,\]

Given a temperature and salinity, Frankenstein and Garner (1967) evaluate polynomials to fit Assur's brine volume data. Their relations are reproduced as follows:

\[-2.06 ^\circ C \leq T < -0.50 ^\circ C \quad V_b = S_b \left( 2.28 + \frac{52.56}{T} \right) 10^{-3},\]
\[-8.20 ^\circ C \leq T < -2.06 ^\circ C \quad V_b = S_b \left( 0.93 - \frac{45.917}{T} \right) 10^{-3},\]
\[-22.9 ^\circ C \leq T < -8.20 ^\circ C \quad V_b = S_b \left( 1.189 - \frac{43.795}{T} \right) 10^{-3},\]
\[-37.8 ^\circ C \leq T < -22.9 ^\circ C \quad V_b = S_b \left( 22.8478 + \frac{3.07984 \times 10^3}{T} + \frac{1.58402 \times 10^5}{T^2} \right.\]
\[\left. + \frac{3.61615 \times 10^6}{T^3} + \frac{3.12862 \times 10^7}{T^4} \right) 10^{-3},\]
\[-43.2 ^\circ C \leq T < -37.8 ^\circ C \quad V_b = S_b \left( 14.145 + \frac{1.6426 \times 10^{-3}}{T} + \frac{6.4947 \times 10^4}{T^2} \right.\]
\[\left. + \frac{8.3945 \times 10^5}{T^3} \right) 10^{-3}.\]

Both these sets of equations enable calculation of $V_b$ given a salinity and temperature, and assuming a fixed density of 926 kg m$^{-3}$. To determine the true brine volume for sea ice having a bulk density of $\rho_{si}$, the calculated brine volume should be multiplied by $\rho_{si}/926$. The corrected volume of brine may then be used in subsequent empirical formulae derived by Vant et al. (1974) for the combination of pure ice and brine in sea ice dielectric mixtures.
3.6.3 Sea ice mixture formulas

Hoekstra and Cappilino (1971) have shown by combining the brine model of Stogryn (1971) and the dielectric mixing formulas of De Loor (1968), that the relative permittivity $\epsilon'_{si}$ of sea ice can be calculated by

$$\epsilon'_{si} = \frac{\epsilon'_i}{(1-3V_b)}$$  \hspace{1cm} (3.26)

where $\epsilon'_i$ and $V_b$ are the dielectric constant of pure ice and the relative brine volume respectively. Vant et al. (1974) indicate using samples results obtained from Bering Sea ice that $\epsilon'_{si}$ is indeed linearly related to $1/(1-3V_b)$ and varies with ice type, but that the constant of proportionality is not exactly $\epsilon'_i$. They obtain different correlation coefficients when comparing the model results with field observations, and find that the highest correlation exists in the case of frazil ice. This is expected since the model of Hoekstra and Cappilino assumes spherical brine inclusions. In the case of columnar ice, brine inclusions are not normally spherical and instead have a cylindrical shape. The shape of brine inclusions in multiyear ice is also not precisely known, and so another model needs to be developed for these two types of ice.

Vant et al. find through experiment that frazil ice generally produces higher losses than columnar ice for similar brine volumes. This may be explained by the different shape and orientation of the brine pockets with respect to the electric field. These problems, combined with the fact that multiyear ice contains a large number of air bubbles (due to brine drainage and extensive recrystallisation) necessitates both consideration of the orientation of brine pockets and the presence of air bubbles. The density of the ice $\rho_{si}$ can be used to determine the volume fraction of air present, since salts comprise a negligible proportion of volume (normally less than 1%). The following equation may be used to find this fraction;

$$V_{air} = 1 - \frac{\rho_{si}}{\rho_i}$$  \hspace{1cm} (3.27)

where $\rho_i = 918$ kg m$^{-3}$. In the same manner as with dry snow, Wiener’s (1910) mixing formula may be used to derive a new model incorporating ‘form’, and a third dielectric, air. Using this technique Vant et al. (1974) derive the following equation for a mixture of ice, brine and air;

$$\frac{\epsilon'_{si} - 1}{\epsilon'_{si} + u} = (1 - V_b - V_{air})\frac{\epsilon'_i - 1}{\epsilon'_i + u} + V_b \left(\frac{\epsilon'_b - 1}{\epsilon'_b + u}\right) + V_3 \left(\frac{\epsilon'_{air} - 1}{\epsilon'_{air} + u}\right)$$  \hspace{1cm} (3.28)

where $\epsilon'_{si}$ is the relative permittivity of the mixture, $V_b$ the volumetric fraction of brine, $u$ the ‘formzahl’ or form number, $\epsilon'_b$ the relative permittivity of brine, $V_3$ is the volume fraction of air, and $\epsilon'_{air}$ is the relative permittivity of air. A value of 40 is assigned for the relative permittivity of air (from Stogryn, 1971). As the relative permittivity of air is 1 the third
term on the right of equation 3.28 becomes zero and is omitted from the calculations. Vant et al. (1974,1978) find through observations that $\varepsilon''_{si}$ is strongly dependent upon salinity and temperature, and to a lesser extent, within first year ice, on ice type and density. They derive an empirical relationship for the dielectric loss ($\varepsilon''_{si}$) of sea ice, after finding that the loss bears a direct relationship with $V_b$. The relationship is dependent on ice type, since salinity varies with age:

$$\varepsilon''_{si} = a + bV_b.$$ \[3.29\]

The constants $a$ and $b$ for three different types of ice are as follows:

- **FRAZIL**  
  $a = 0.0316$  
  $b = 9.26$

- **COLUMNAR**  
  $a = 0.0040$  
  $b = 7.143$

- **MULTIYEAR**  
  $a = 0.0063$  
  $b = 9.98$

Vant et al. observe that there is a high correlation between observed results and modelled results for $\varepsilon''_{si}$ in both frazil and multiyear ice when a high form value ($\infty$) is substituted in equation 3.28. A low correlation is observed in the case of columnar samples because brine pockets were orientated perpendicular to the incident electric field, and thus had less effect than if parallel. Results demonstrate that a model of this sort is effective with a high form factor, but that equations which include a ‘form factor’ have drawbacks. They have the effect of masking to some degree the behaviour that one is trying to model, by lumping the shape and orientation of brine pockets into a semi-empirical ‘formzahl’. Nonetheless, such a simple model is preferred to a more complicated theoretical formula, as ice conditions, brine inclusion shape, size, and orientation are not known accurately enough for such factors to be incorporated. Even in a reasonably realistic model, it would appear that at GHz frequencies it is impossible to include all effects of impurities in the ice. Vant (1976) suggests that the effects of scattering by air bubbles, for example, may only be completely excluded below a frequency of 4 GHz for first year ice, and 1.5 GHz for multiyear ice. Thus, at a frequency of 10 GHz or more, values for $\varepsilon''_{si}$ are likely to be higher than those predicted by the formulae, because of the effects scattering will have. So, these formulae are considered a useful first order calculation of the dielectric constants and losses of the sea ice.

Here the formulae used incorporate the effects of increasing salinity and density, and varying temperature, so that their combined effects upon the electrical properties of particular sea ice/brine mixtures may be investigated. The dielectric and reflective properties of sea ice have been calculated using equations 3.27 and 3.29 for a variety of ice densities, form factors, and salinities, but all at a fixed frequency of 10 GHz. This is a restriction imposed by the empirical equations of Vant et al. (1974), whose work was undertaken at selective
wavelengths only. First of all an salinity of $5 \, ^{\circ}/_{\infty}$ was chosen (marking the upper limit to salinity observations in Figure 3.10) to investigate the effects of varying form factor for both frazil and columnar ice at a fixed density of 850 kg m$^{-3}$ for comparison. The results are plotted in Figure 3.11a and 3.11b respectively. As temperature increases the brine volume increases, but with increasing rapidity just below 0°C. The rapid increase in both relative permittivity and dielectric loss at temperatures close to melt point is caused by the brine volume relationship originally found by Assur (1960), and demonstrated by Frankenstein and Garner (1967) in the earlier equations. The effect which rapidly increasing brine volume has between -10°C and 0°C is to cause a proportionate increase in both the real and imaginary parts of the complex dielectric constant. In fact, between -2°C and 0°C both values rapidly tend to infinity, for a large form factor (ie. $f = \infty$). Vant et al. (1974) suggest from their experiments that $f = \infty$ is most applicable to the three classes of ice under scrutiny, and so for frazil at -1°C values of over 7.5 and 1.2 occur for $\epsilon'$ and $\epsilon''$ respectively. The range of temperatures experienced during the course of surface observations from FS Polarstern (Burns et al., 1986) are indicated to highlight the bounds for values of the dielectric properties of these types of sea ice. For a given salinity the dielectric losses are less in columnar ice than in frazil, this being evident in the values of $\epsilon''$ in Figures 3.11a and 3.11b. This may be an artefact of the way in which their samples were oriented when dielectric measurements were originally made, and so it is safest for these purposes to treat frazil and columnar ice as one category called ‘first year ice’. The orientation of brine pockets, or a high loss dielectric within an ice matrix has a large role to play in the dielectric constant. A high form factor assumes that the long axes of all brine pockets are oriented roughly vertically in the ice matrix (see Figure 3.3). This is a fairly simple assumption to make, yet it enables the distinction of an upper limit to the magnitude of the dielectric constant. In the case of multiyear ice, densities and salinities are normally lower than first year ice. Figure 3.11c shows that values for $\epsilon'$ and $\epsilon''$ fall by 50% and 75% for a given temperature at typical values of $1 \, ^{\circ}/_{\infty}$ salinity and density 700 kg m$^{-3}$. At lower salinities the effects of increasing brine volume and temperature are less marked, and the increase in $\epsilon'$ between -40°C and 0°C when $u = \infty$ is only 1.0.
Figure 3.11  Sea ice dielectric mixture formula results using different Formzahls. $\epsilon'$ and $\epsilon''$ are plotted at varying temperatures for (a) frazil ice of 5 °/oo salinity and a density of 850 kg m$^{-3}$, (b) columnar ice of 5 °/oo salinity and a density of 850 kg m$^{-3}$, and (c) multiyear ice of 1 °/oo salinity and a density of 700 kg m$^{-3}$. 
3.6.4 Reflexion coefficient of sea ice

Providing a high form factor is used (Vant et al., 1974) then the variability in the reflexion coefficient \( (R) \) may be investigated for these three different types of ice using different values of temperature and salinity (with density constant), or temperature and density (with salinity constant). In Figure 3.12a \( R \) is plotted at different temperatures for frazil ice of density 850 kg m\(^{-3}\). A family of curves is drawn indicating the relative increase in the reflexion coefficient for different salinities. At a salinity of 4 ppt, \( R \) rises from a minimum of -11.5 dB at temperatures below -40°C, and attains a maximum of over -6 dB at -2°C (tending to 0 dB at 0°C). At the uppermost limit of salinity observed in frazil ice of 16 ppt, the minimum of around -11 dB at -40°C is less than 1 dB higher than for lower salinities. The difference in \( R \) is greatest between -5 and -20°C for different salinities, but near melting point \( R \) tends rapidly to unity in all cases.

![Figure 3.12](image)

**Figure 3.12** Effect of varying temperature and salinity upon the reflexion coefficient of (a) first year sea ice of density 850 kg m\(^{-3}\) and (b) multiyear sea ice of density 750 kg m\(^{-3}\).

Upon investigation of the reflective properties of columnar ice at similar salinities, only the dielectric losses are observed to change when different constants \( a \) and \( b \) are substituted in equation 3.29. Otherwise the relative permittivity calculated using the empirical relationship in 3.28 stays the same. Despite dielectric losses being reduced, such small changes result in minimal effect upon \( R \) at similar densities and salinities. Maximum observed differences are a fraction of a dB only. This justifies a joint classification of frazil and columnar ice types into a first year ice category, on the grounds of their similar reflective properties.

In Figure 3.12b the same technique is used to investigate the reflective properties of
multiyear ice of density 750 kg m\(^{-3}\). In contrast, \( R \) rarely exceeds -10.5 dB except above \(-2^\circ C\), and falls to a minimum of around -12 dB below \(-20^\circ C\). Maximum variability in \( R \) is only 2 dB throughout the whole range of temperatures and salinities likely to be experienced during MIZEX '84.

**Figure 3.13** Effects of varying temperature and salinity upon the reflection coefficient of (a) first year ice of salinity 16 ppt, and (b) multiyear ice of salinity 1 ppt.

Finally, Figures 3.13a and 3.13b demonstrate the effects of varying the density of first year and multiyear ice. In the case of first year ice (Figure 3.13a) curves are plotted for different temperatures and densities for a high salinity. The variation in \( R \) is most marked at lower temperatures, where differences of up to 2 dB are observed for the typical range of densities encountered. Figure 3.13b shows a similar effect for multiyear ice (of 1 °/\( \infty \) salinity) at different temperatures for a different range of densities. Variability in \( R \) with temperature is minimal below \(-5^\circ C\), but the relative differences in \( R \) with density around \( 0^\circ C \) are much greater than for higher salinity first year ice.
3.7 Summary

Analysis of the dielectric and reflective properties of different snow and ice media has demonstrated the extreme variability in electromagnetic properties for seemingly small changes in material properties. The comparative evaluation of these findings leads to the conclusion that the dielectrics of the upper layers of terrestrial ice and sea ice may differ substantially enough to be able to detect these on the basis of their coefficient of reflection. The response of an active radar such as a radar altimeter will thus depend upon the recent climatic conditions, and other factors influencing the combinations of dielectrics of polar surface materials. The main results from this analysis are:

- A documentation of the range of dielectric properties of ice and snow under different temperature regimes, and water and impurity contents.
- Determination of the reflection coefficient at normal incidence from an air/snow or air/ice dielectric interface, under a range of conditions.
- Determination of the loss factor in certain materials, a feature which assumes importance in the subsequent discussion of energy which penetrates the surface. These results are extended to the treatment of the scattering properties of surfaces with differing degrees of roughness in the following chapter. This is the next logical step in understanding the response of the altimeter in different circumstances.
CHAPTER 4
REFLEXION AND SCATTERING FROM ROUGH GLACIERIZED TERRAIN

Fresnel reflexion, governed by the electromagnetic properties of snow or ice media, and described in Chapter 3, only applies to perfectly smooth planar surfaces bounding homogeneous media. The results of the previous chapter cannot therefore, in general, be applied directly to the real world, because of irregularities or terrain of snow and ice surfaces. Nonetheless, the fundamental concepts outlined in the previous chapter are a necessary stage before understanding the scattering response of any surface.

In this chapter the scattering response from a variety of different terrestrial ice and sea ice surfaces surfaces is examined using theoretical and empirical models. Rough surface scattering is treated and the fundamentals introduced for surface reflexion are extended to internal scattering from dielectric discontinuities occurring in layered materials. The aim of this chapter is to characterise the scattering signatures of various surface media in order that differences in scattering signatures may be recognised in RAL radar altimeter data as a response to different material properties or different surface roughnesses.

4.1 The Rayleigh criterion

The first problem that arises is to decide whether snow or ice surfaces may be considered smooth or whether irregularities or perturbations from a plane are large enough for the surface to be considered rough electromagnetically. This distinction is critical in terms of the proportion of incident energy which is returned to a radar altimeter instrument. If on the one hand the surface is smooth then energy is either reflected back or transmitted into the snow or ice medium; but if on the other hand the surface is rough then incident energy may be scattered in many directions.

Rayleigh proposed that the surface be considered smooth if the phase difference $\Delta \Phi$ between waves reflected from the foot and apex of an irregularity on the surface is less than $\pi/2$. Following Beckmann and Spizzichino (1963) Figure 4.1 shows that in the case of a surface with irregularities of maximum height difference $\Delta h$, waves reflected at $A$ and $B$ will be shifted in phase with respect to each other by:

$$\Delta \Phi = \frac{4\pi \Delta h \cos \theta}{\lambda} \quad \{4.1\}$$

where $\lambda$ is the wavelength and $\theta$ is the incidence angle. Thus, for a surface to be smooth, the
Figure 4.1  Path difference between fields of rays reflected from points A and B with a height difference $\Delta h$ is $\Delta r$, where $\Delta r = 2\Delta h \cos \theta$. The phase difference $\Delta \Phi$ is $\Delta \Phi = 2k\Delta h \cos \theta$ and $k = 2\pi/\lambda$.

Rayleigh criterion

$$\Delta \Phi < \frac{\pi}{2}$$  \hspace{1cm} \{4.2\}

becomes

$$\Delta h < \frac{\lambda}{8 \cos \theta}.$$  \hspace{1cm} \{4.3\}

For a random surface characterised by a standard deviation of surface height ($\sigma_h$) the $\Delta h$ is interchangeable with $\sigma_h$ (Ulaby et al., 1982). For the RAL altimeter this means that surfaces having a small-scale roughness component with $\sigma_h$ of the order of 3 mm or more may be considered rough.

The Rayleigh criterion supplies a qualitative indication of roughness, using an order of magnitude approach to define the dividing line between rough and smooth. However, according to some authors (Kerr, 1951; and Burrows and Attwood, 1949) the $\Delta h$ is not defined precisely enough. They suggest that the coefficient 1/8 in equation 4.3 is often better replaced by values 1/16 or 1/32, for example. Ulaby et al. (1982) also maintain that the Rayleigh criterion is only useful as a first-order classifier of surface roughness. In cases where the wavelength is of the order of $\sigma_h$ a more stringent criterion is suggested necessary: it is named the Fraunhofer criterion. For a surface to be considered smooth this condition requires that the maximum phase difference between rays received at the centre and edge of the altimeter antenna pattern are less than $\pi/8$ radians. This reduces to:

$$\sigma_h < \frac{\lambda}{32 \cos \theta}.$$  \hspace{1cm} \{4.4\}

This criterion is more stringent than the Rayleigh criterion, and furthermore is consistent with experimental observations. Applying such a rule to the broad-beam RAL altimeter, where the maximum incidence angle at the edge of the 3 dB beam-limited footprint is $5^\circ$, $\sigma_h$ must be less than 0.69 mm for the surface to appear smooth. In reality, however, there is no dividing line
between smooth or rough conditions, and so there is little sense in trying to find a more exact definition than the Fraunhofer criterion. It would seem that under almost all circumstances snow and ice surfaces may be assumed 'rough' at the millimetre scale.

4.2 Effects of surface roughness upon scattering characteristics

Qualitatively speaking, the term 'smooth surface' is used to describe a surface that obeys the Fresnel reflexion and transmission laws discussed in the previous chapter. Experimentally a surface may be considered smooth if the observed scattering coefficient is in close agreement with values calculated when the surface is assumed a perfect specular reflector.

![Diagram of scattering patterns and angular variation in backscatter coefficient](image)

**Figure 4.2** (a) Scattering patterns for surfaces of three different degrees of roughness. Surface roughness is schematic and not to scale. (b) Angular variation in the backscatter coefficient $\sigma^0$ for corresponding roughnesses.

As the surface is transformed from a specular to slightly rough surface the characteristic delta function shape for a smooth surface changes into a narrow lobe (see Figure 4.2(a)), the
axis of which is oriented along the specular direction; and in addition the pattern includes a broad beam lobe of lower magnitude due to diffuse scattering. Figure 4.2(b) schematically illustrates the variation in backscatter coefficient \((\sigma^o(\theta))\) with angle of incidence for surfaces of differing roughness. At nadir and normal incidence for a smooth surface, the backscatter as a function of \(\theta\) for the specular surface is a delta function, due to phase coherent addition of returns. As the roughness of the surface increases, more returns from individual point scatterers become out of phase, and the magnitude of the coherent nadir component is decreased. Then \(\sigma^o(\theta)\) assumes the shape of an exponentially decaying function with \(\theta\), for the region close to nadir. That is, at near normal incidence the following condition is satisfied (Ulaby et al., 1982),

\[
\sigma^o(\theta) = \sigma^o(0)e^{-a\theta}
\]

{4.5}

where \(a\) is a constant. Because \(\sigma^o(\theta)\) may decrease by two or more orders of magnitude, as a function of increasing incidence angle, it is normal practice to express and plot \(\sigma^o\) in dB to compress its scale. Thus,

\[
\sigma^o(\theta) = 10\log_{10} \sigma^o(\theta)
\]

{4.6}

and for the exponential form given by 4.5, this is expressed as

\[
\sigma^o = \sigma^o(0) - a'\theta
\]

{4.7}

where \(a' = 4.34\ a\), and its units are in dB per degree with \(\theta\) in degrees. Hereafter it is assumed that \(\sigma^o\) is expressed in dB.

From a measurement point of view, Ulaby et al. (1982) suggest that a surface may be considered specular if the backscattered power from off-nadir angles is so small that it is difficult to measure. Therefore, the surface may be considered smooth if for example \(\sigma^o\) decreases by more than 40 dB between \(\theta = 0^\circ\) and \(10^\circ\), which corresponds to \(a' \geq 4\) dB/degree. In previous literature, discussions related to the backscatter coefficient of a ‘smooth surface’ usually interpret it as meaning merely that \(\sigma^o(\theta)\) decays rapidly with increasing \(\theta\), and not necessarily that it is composed entirely of a true specular component (ie. perfect plane reflexion) in the immediate vicinity of the nadir direction. From this point onwards, therefore, the scattering from relatively smooth surfaces will be termed ‘quasi-specular’, since it is seldom the case that the Rayleigh criterion is met for an electromagnetically smooth surface.

Regardless of whether one is interested in the theoretical or experimental aspects of scattering or in the practical applications of the problem, the study of scattering response always brings to light two fundamentally different phenomena; specular reflexion and diffuse scattering (Beckmann and Spizzichino, 1963). Scattering from typical glacierized terrain can
be resolved into two components; coherent or specular returns, and non-coherent or diffuse returns. The terms coherent and non-coherent are sometimes used in subsequent sections to help describe the scattering phenomenon. If the phase of a scattered wave is constant or varies in a deterministic manner, it is coherent. If the phase of a wave is random and uniformly distributed over an interval of phase $2\pi$, it is non-coherent (Long, 1983). Most importantly, for $n$ waves of equal power, the total power is directly proportional to $n$ if the waves are non-coherent and to $n^2$ if the waves are phase-coherent. In circumstances where plane reflexion occurs the coherent component will therefore dominate over the non-coherent or diffuse component of scattering.

4.3 Surface scattering

Microwave energy incident upon a rough surface is partly reflected in the specular direction and partly scattered in all directions. A monostatic radar (ie. transmitter and receiver at the same locations) receives the backscattered component of the scattered energy (ie. energy returned directly back to the receiver). As the surface becomes rougher the coherent component becomes negligible and the non-coherent component is reduced. For very rough surfaces the scattering pattern approaches that of a Lambertian surface, or perfect isotropic scatterer. The scattering characteristics of three types of surface are illustrated in Figure 4.2 for normally incident radiation.

4.3.1 Simple physical models

Later in this chapter a more detailed mathematical approach is used to the scattering problem, but an understanding of many of the physical scattering mechanisms can be based on relatively simple concepts. Ulaby et al. (1982) discuss two simple models which are appropriate and applicable in later sections:

Point scatterers

Some of the earliest models assume that surface scattering results from many individual point scatterers, each of which is isotropic. This is, however, suitable only for surfaces having scattering components whose heights above or below the mean surface level is small compared to a wavelength. Clapp (1946) was one of the writers to develop several models of this kind. However, the assumptions necessary make such a model inadequate for most polar surfaces.

Plane Facets

An alternative approach to that of Clapp involves the use of ‘facet models’ where the surface is approximated by an assemblage of facets tangential to the actual surface. Facet
models treat scattering by taking into account the distribution of slope facets and the re-radiation pattern of reflected energy. Scattering from each facet depends on its size relative to the wavelength. If the facet is wide (i.e. many wavelengths across) then scattering is by specular reflexion. However, the only way radiation may be returned to the point source is from large facets oriented normal to incident energy. Thus for the purposes of modelling, facets are considered infinite in extent (relative to a wavelength). For this condition to be satisfied under all circumstances the wavelength is assumed zero. Such an assumption is implicit in geometric optics, so that these facet models have become known as 'geometric-optics scattering models'.

4.3.2 Modification of the reflexion coefficient by roughness

If, as various theoretical and experimental observations have indicated, the total scattered field by a rough surface can be considered to be the sum of a specular and a diffuse component, then we must consider the effects of roughness upon the reflexion coefficient and these components. According to the Rayleigh roughness criterion, effective roughness depends upon the surface height irregularities and the incidence angle, for a given wavelength. The apparent reflexion coefficient, in the same manner, is also a function of these parameters. The determinants of the fraction of energy backscattered from the surface are the reflexion coefficient of a smooth surface \( R_o \) and a surface roughness factor \( \rho \). The following expression is used by Ament (1953), Beckmann and Spizzichino (1963), and Long (1983), to express the magnitude of the reflexion coefficient \( R_{\text{rough}} \) for a rough surface:

\[
R_{\text{rough}} = \rho R_o
\]

where \( \rho \) varies between 0 and 1 depending upon whether the surface is extremely rough or perfectly smooth, respectively. Various theoretical expressions have been derived for \( \rho \) which correspond to different models of surface roughness. Beckmann and Spizzichino (1963) use the following expression for \( \rho \) for a Gaussian distributed surface:

\[
\rho = \sqrt{e^{-(\Delta \Phi)^2}}
\]

where the phase difference \( \Delta \Phi \) is calculated from equation 4.1. From experimental data Beckmann and Spizzichino (1963) postulate that a phase difference exceeding 0.4\( \pi \) probably corresponds either to diffuse scattering or to a combination of specular or diffuse. Thus for definition of terms the roughness factor may be referred to as \( \rho_s \) or \( \rho_d \), in the case of specular or diffuse scattering respectively. Then the reflexion coefficient of the diffusely scattered component, for instance, is given by

\[
R_{\text{rough}} = \rho_d R_o.
\]
Beckmann and Spizzichino review the results of experiments to determine the typical value for $\rho_d$ under a wide variety of conditions, and conclude that most of the results for the roughness factor suggest a value of between 0.45 and 0.63, with a mean at around 0.57. Also power contained in the incoherent field was found to contribute about 30%, at most, of that in the coherent field.

Assuming a value of 1 for $R_o$, as with a smooth perfect specular surface, then $R_{\text{rough}} = \rho_s$. The mean reflexion coefficient $R_{\text{rough}}$ for a surface with Gaussian distributed surface heights may therefore be calculated using equations 4.1 and 4.9. In Figure 4.3 the results of calculating the value of $R_{\text{rough}}$ are plotted for a surface of varying roughness. The coefficient of reflexion is reduced significantly as soon as the Fraunhofer criterion is met (i.e. when $\rho = 0.92$). The experimentally derived mean value of $\rho_d = 0.57$ is included to define the regimes in which specular and diffuse scatter may occur. Over the range of incidence angles shown there is little change in the value of $R_{\text{rough}}$ at any of the roughnesses indicated.

![Figure 4.3](image_url)

**Figure 4.3** Variation in mean reflexion coefficient for a perfectly reflecting Gaussian distributed surface (thus $R_o = 1$), at a frequency of 13.8 GHz.

### 4.3.3 The Kirchoff or physical optics scattering model

Among the many theories for surface scattering, the Kirchoff or physical optics formulation is one of the most widely used (Fung and Chan, 1969; Fung, 1981; and Fung and Eom, 1982). The theory is applicable to surfaces with undulations whose average dimensions are large compared with the incident wavelength. Its tacit assumption, as with the facet model previously discussed, is that plane-boundary reflexion occurs at every point on the surface. That is, at any point of incidence the total field can be computed as though the incident wave
impinges upon an infinite plane tangential to the surface. Wu and Fung (1972) call this a 'stationary-phase approximation', since backscattering can only take place by phase-coherent reflexion from points at which the surface is normal to the incident waves. This case of backscattering in surface scattering is of special interest, since radar altimeter measurements fall into this category. For surfaces with a large standard deviation of surface heights Ulaby et al. (1982) derive the following equation for the backscatter coefficient, using a stationary-phase solution:

\[ \sigma^2(\theta) = \frac{|R(0)|^2 \exp\left(-\tan^2 \theta / 2\delta^2\right)}{2\beta^2 \cos^4 \theta} \]  \tag{4.11}

where \( R(0) \) is the Fresnel reflexion coefficient evaluated at normal incidence from equation 3.3 in chapter 3, and \( \beta \) is the rms surface slope (in radians). The main assumption of equation 4.11 is that the surface is random and has a Gaussian distribution of heights with probability

\[ p(h) = (2\pi \sigma_h^2)^{-0.5} \exp\left(-\frac{h^2}{2\sigma_h^2}\right), \]  \tag{4.12}

when \( h \) is the height above the mean surface, and \( \sigma_h^2 \) is the variance of surface heights. Ulaby et al. (1982) specify that certain conditions must be met regarding the standard deviation of surface heights (\( \sigma_h \)) and the surface correlation length (\( l \)) of surface perturbations: the correlation length is simply the distance between two points on the surface (or spatial displacement) for which the normalised autocorrelation function is equal to 1/e. They suggest that these two parameters, when combined, may be used to describe the statistical variation of the random component of surface height. Thus when statistical surfaces are considered using this technique, their correlation length \( l \) or horizontal scale roughness, must be larger than the wavelength, while their standard deviation of surface heights, or vertical scale roughness must be small enough so that the average radius of curvature is larger than a wavelength. The above-stated restrictions may be expressed mathematically as,

\[ kl > 6.0 \]  \tag{4.13}

\[ l^2 > 2.76 \sigma \lambda \]  \tag{4.14}

\[ k\sigma > 2.0 \]  \tag{4.15}

where \( k \) is the wavenumber or \( 2\pi/\lambda \).

For the use of equation 4.11, surface slopes may be chosen to be small, but the surface itself can never approach a plane. The scattering coefficient is thus applicable to many rough surfaces encountered by an altimeter. However, it should be recalled that this type of surface scattering model does not take into account multiple scattering or shadowing, and for such situations modifications may be necessary (Fung and Eom, 1979). When the criteria for its
impinges upon an infinite plane tangential to the surface. Wu and Fung (1972) call this a 'stationary-phase approximation', since backscattering can only take place by phase-coherent reflection from points at which the surface is normal to the incident waves. This case of backscattering in surface scattering is of special interest, since radar altimeter measurements fall into this category. For surfaces with a large standard deviation of surface heights Ulaby et al. (1982) derive the following equation for the backscatter coefficient, using a stationary-phase solution:

\[
\sigma^0(\theta) = \frac{|R(0)|^2 \exp(-\tan^2\theta/2\beta^2)}{2\beta^2 \cos^4 \theta} \tag{4.11}
\]

where \( R(0) \) is the Fresnel reflection coefficient evaluated at normal incidence from equation 3.3 in chapter 3, and \( \beta \) is the rms surface slope (in radians). The main assumption of equation 4.11 is that the surface is random and has a Gaussian distribution of heights with probability

\[
p(h) = (2\pi \sigma_h^2)^{-0.5} \exp \left( \frac{-h^2}{2\sigma_h^2} \right), \tag{4.12}
\]

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use are met, surface scattering is purely non-coherent and is a function of the incidence angle ($\theta$). Figure 4.4 shows the backscattering characteristics of five surfaces with different rms slopes using equation 4.11. The backscatter coefficient $\sigma^0$ is normalised with respect to the Fresnel reflection coefficient $R$ at normal incidence so that the curves are applicable to any Gaussian distributed surface of known reflection coefficient. As expected, $\sigma^0$ falls less rapidly with increasing $\theta$ as the rms slope $\beta$ (radians) increases. For surfaces with small rms slopes, for instance, a strong peak occurs around normal incidence, but it falls off rapidly between 0 and 10°: this shows some similarity to the situation in Figure 4.2(b). At low rms slopes the surface thus has a large coherent component of scatter. Such scattering is referred to as 'quasi-specular' in the context of altimeter returns demonstrating these characteristics.

![Figure 4.4](image)

**Figure 4.4** Normalised backscatter coefficient ($\sigma^0/R$) calculated for varying rms surface slope using a Kirchhoff approximation with a Gaussian distribution of surface heights.

Most naturally occurring surfaces are not spatially geometrically homogeneous, or physically homogeneous, and few contain only one scale of roughness (Long, 1983). Care must be exercised therefore to ensure that backscatter data samples which are to be compared are, to the best of knowledge, collected from statistically homogeneous areas. This means, for example, that the dielectric properties of the medium and the scattering characteristics of each area should as far as possible be constant.

Under natural circumstances surface scattering may not be the only mechanism contributing to the coefficient of backscatter $\sigma^0$. Rarely, if ever, is the coefficient of reflection unity, and so a large proportion of incident radiation is transmitted into the medium. Providing attenuation effects within the surface layer medium are low, volume scattering from
inhomogeneities within the layer may contribute to $\sigma^\circ$. Thus, only in cases where penetration and volumetric contributions are negligible, can one apply a surface scattering model alone. There is one exclusive situation in polar regions when this occurs, and it is for open ocean surfaces. Since brine has a relative permittivity and dielectric loss of about 30 and 38, respectively, penetration is effectively negligible. In such a case volumetric scattering effects may be ruled out.

### 4.3.4 Ocean surface scattering

The phenomena involved in radar backscattering from the sea are reasonably well understood (Katzin, 1957; Long, 1983; and Chaudhry and Moore, 1984). As explained, sea water is a highly conductive or high loss medium, and so scattering takes place at the surface only. Backscattering is strongly related to the wind and wave fields through the slope distribution of the surface. Theoretical expressions have been developed to relate $\sigma^\circ$ to the mean square surface slope by Barrick (1974), and Hammond et al. (1977). Foundations for these models lie in the Kirchoff or physical optics approach used in the previous section, and the expressions derived are almost identical. Results for the scattering response of a Gaussian distributed ocean surface may therefore be calculated using equation 4.11. or from Figure 4.4 using experimental values for the reflection coefficient at normal incidence $R(0)$. Robinson (1985) gives values of between -2.08 and -2.37 dB for $R(0)$ at normal incidence for sea water.

Ocean surfaces may be described in terms of the spectrum of their vertical displacements from the mean surface. Under conditions of a fully developed sea, the full spectrum of waveheights and wavelengths are present. Waves present many facets which reflect incident energy in different directions, since surface slopes may be tilted up to 20 or even 25° from the horizontal. In this case the sea surface acts like a Lambertian or isotropic surface, giving equal scatter in all directions. Figure 4.4 shows that this situation is reached for a Gaussian distributed random surface with an rms slope of 0.45 or more. For a nadir viewing altimeter this means that ocean backscatter under the majority of circumstances is non-coherent, or diffuse, with no strong coherent component present at near nadir incidence angles. Although diffuse ocean scatter may be regarded as the norm, there are two special cases in which ocean surfaces can cause quasi-specularity. The first of these is when the ocean surface is flat calm. This situation requires either no wind, the growth of grease ice, or wave suppression by damping by sea ice floes. In the second, phase-coherent returns are possible from rough surfaces when the wavelength of the incident pulse is resonant with components of the ocean surface.

A phase-coherent sum of reflections from individual surface perturbations may occur only when the surface roughness has a particular wavelength. This phenomenon is described by the term Bragg resonance (Ulaby et al., 1982), by analogy with the Bragg resonances used in
spectroscopy. Figure 4.5 illustrates the phenomenon. The incidence angle of the plane wave is $\theta_i$ and the wavelength of the surface component is $L$, while the radar wavelength in this case is $\lambda$ (0.022 m). If the excess distance $\Delta R$ from the source to each successive crest is $\lambda/2$, the roundtrip phase difference is $\lambda$ and the signals add in phase. If $\Delta R$ is any other distance then they add out of phase and signal strength is reduced (owing to non-coherent returns).

![Diagram](image)

**Figure 4.5**  A schematic diagram of the conditions required for Bragg resonance, simplified for a large angle of incidence $\theta_i$. Coherent backscatter is given by waves whose crest surfaces are perpendicular to incident energy, and whose wavelength $L$ is given by $L = \Delta R/(2 \sin \theta_i)$.

The Bragg resonance condition is written in the form:

$$\frac{2L \sin \theta_i}{\lambda} = n \quad \text{where} \quad n = 0, 1, 2 \ldots \quad \{4.16\}$$

Kwoh and Lake (1984) observe that the wavelength is resonant to components of the surface that are either very short gravity waves, or capillary or surface tension waves (Phillips, 1969). The former may reach sizes which are of importance to scattering, and have a minimum wavelength governed by the relation, (Phillips, 1969);

$$\lambda > 2\pi \left(\frac{\gamma}{g}\right)^{1/2}, \quad \{4.17\}$$

where $\gamma$ is the surface tension and $g$ is the acceleration of gravity. In clean sea water at 20°C, the minimum phase velocity for gravity waves is 0.23 m s$^{-1}$, and the shortest wavelength is approximately 1.7 cm. The size of such short gravity waves satisfies the Rayleigh criterion of roughness and enables the resonant mechanism of Bragg scattering to take place. However, in shorter waves the restoring force per unit displacement is predominantly the result of surface tension, and so such waves are usually called capillary waves. Crapper (1957) finds the limiting range of amplitude for capillary waves and gives a family of solutions for steady waves of finite amplitude. Typically these very short ocean waves ride upon larger ocean waves, yet the Bragg resonance effect is so strong that capillary waves of the order of 1 mm high can be a
dominant scattering mechanism, even when the underlying ocean waves are many metres in height. Capillary waves form just in front of primary wave crests in waves where the primary wavelength is short, such as wind generated storm waves. They are parasitic, acquiring their energy by extraction from the wave crests, and do not form as a result of wind action as do short gravity waves.

Table 4.1 Table of values of ocean wavelength components resonant with a 0.022 m wavelength (ie. 13.8 GHz) at varying local incidence angles.

<table>
<thead>
<tr>
<th>Local Incidence Angle (Degrees)</th>
<th>Resonant Wavelength ($\times 10^{-2}$ m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>63.0</td>
</tr>
<tr>
<td>3</td>
<td>21.0</td>
</tr>
<tr>
<td>5</td>
<td>12.6</td>
</tr>
<tr>
<td>10</td>
<td>6.30</td>
</tr>
<tr>
<td>20</td>
<td>3.20</td>
</tr>
<tr>
<td>40</td>
<td>1.71</td>
</tr>
<tr>
<td>60</td>
<td>1.27</td>
</tr>
</tbody>
</table>

Moore (1981) gives an idea of the range of surface wave components that are resonant at different microwave frequencies and different incidence angles. Table 4.1 is a similar range of values calculated from equation 4.16 at a wavelength of 0.022 m. This table indicates that true Bragg resonance for capillary waves is only important for angles of incidence > 20°. At near-normal incidence longer waves may contribute to resonant returns, but these components of the ocean surface (larger than the wavelength of incident energy) are not observed to dominate backscatter to the same degree.

Although Bragg scattering only occurs for large local incidence angles this type of phase-coherent return is not excluded for an altimeter. As capillaries and short gravity waves normally ride on the backs of longer waves or swell, the scattering mechanism is modulated by the effects of the long wavelength components. The principal cause of this mechanism known as ‘tilt modulation’ is the way in which the slope of the underlying long waves presents the ripples to pulses transmitted from the radar. The combined effects of an incidence angle of 5°, at the outer limits of the RAL altimeter beam limited footprint, and local tilt of 10° or more by long waves, is sufficient to cause true Bragg resonant effects. Radar backscatter will tend to be strongest from the slope of the long waves facing away from the altimeter because of the combined incidence and tilt angles. The actual magnitude of Bragg scatter in such circumstances depends upon the spectral distribution of short wave energy present, since changing the angle of view of the surface slightly changes the length of ripples which will be
in Bragg resonance. So providing the right roughness components are present an essentially rough ocean surface may produce powerful coherent returns, from the combined effects of a large number of in-phase point scatterers.

### 4.4 Volume scattering

When an electromagnetic wave impinges upon the boundary between air and a snow or ice medium which is effectively semi-infinite, a proportion is backscattered and the rest is transmitted into the medium. As we have seen from the previous section and Chapter 3, surface scatter is determined by a combination of the reflection coefficient and the surface roughness. If the lower medium is inhomogeneous or is a mixture of different dielectrics (as with snow or ice) then some of the energy which is transmitted into the medium may be scattered back into the air, by the inhomogeneities. This process is referred to as volume scattering. The mechanism of volume scattering causes a redistribution of the energy in the transmitted wave into other directions, and results in energy loss from the transmitted wave. In general, a wave propagating within a snow or ice medium will also experience losses due to conduction. The total loss (i.e. sum of scattering and conduction losses) is usually referred to as the extinction, and the extinction per unit length referred to as the extinction coefficient \( k_e \).

![Figure 4.6](image)

**Figure 4.6** Qualitative illustration of the variation in the volume backscatter coefficient \( \sigma_v^0 \) with incidence angle. Large values for the complex dielectric constant cause higher transmission losses and so \( \sigma_v^0 \) is reduced.

To determine the presence of volume scattering it is necessary to know: (a) what is the
effective depth of penetration into the medium, and (b) if the medium may be considered homogeneous. With surface scatter, the magnitude of the backscatter coefficient is proportional to the relative dielectric constant of the surface, while the angular scattering pattern is governed by surface roughness. In comparison, volume scattering strength is proportional to the dielectric discontinuities within the medium, while the angular variation in scattering is determined by the average dielectric constant of the medium, the geometric proportions of the inhomogeneities relative to the incident wavelength, as well as the surface roughness. As an illustration, for an inhomogeneous medium with a small average dielectric constant, the angular backscattering curve should in point of fact be fairly uniform. For a larger average dielectric constant Ulaby et al. (1982) maintain that the volume backscatter coefficient falls more rapidly with increasing incidence angle. A qualitative illustration of volumetric backscatter as a function of incidence angle is shown in Figure 4.6.

4.4.1 Penetration depth

Ulaby et al. (1984) derive an expression for the depth of penetration ($\delta_p$) in a medium. $\delta_p$ defines the depth beneath the surface to a point at which the average power of the propagating wave is equal to $e^{-1}$ of its level at the surface. In general, $\delta_p$ is governed by scattering and absorption losses in the medium. If scattering losses can be ignored then $\delta_p$ is given by:

$$
\delta_p = \frac{\lambda}{4\pi} \left\{ \frac{\epsilon'}{2} \left[ \left( 1 + \left( \frac{\epsilon''}{\epsilon'} \right)^2 \right)^{1/2} - 1 \right] \right\}^{-1/2} \tag{4.18}
$$

where $\lambda$ is the wavelength in air and $\epsilon'' = \epsilon' - j\epsilon''$ is the relative complex dielectric constant. Furthermore, if the loss tangent $\epsilon''/\epsilon' < 0.1$ then the preceding expression can be simplified to:

$$
\delta_p \approx \frac{\lambda\sqrt{\epsilon'}}{2\pi\epsilon''} \tag{4.19}
$$

which is applicable to most natural materials except water (Ulaby et al., 1982).

In Figure 4.7 a, b, c, and d penetration depth curves of typical snow and ice mediums are calculated for a frequency of 13.8 GHz. Theoretically, the penetration depth may be of the order of several metres for a dry snowpack (see Figure 4.7a), although in reality scattering losses within the medium will reduce this value. On the other hand the depth of penetration is reduced to the order of magnitude of a wavelength when the medium has only a small fraction of free water or brine. As an example, at a snow density of 500 kg m$^{-3}$, if we increase the amount of water by volume from 0 to 3% then the penetration depth is reduced significantly from several metres to about 3 cm.
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Figure 4.7 Penetration depth curves for snow and ice media of different composition and impurity content. (a) illustrates the effect of introducing water of known content into dry snow of a given density. (b) shows the variation in $\delta_p$ occurring in frazil ice of density 850 kg m$^{-3}$ for varying temperature and salinity. (c) gives the variation in columnar ice of similar salinity and density, and (d) illustrates changes in $\delta_p$ in less saline multiyear ice of density 750 kg m$^{-3}$. 
For first year ice of relatively high salinity penetration depths are minimal due to the high dielectric loss of brine. There are differences between frazil ice in Figure 4.7(b) and columnar ice in Figure 4.7(c), and the variation in $\delta_p$ with salinity in the latter is much larger - owing to their dissimilar structure and brine pocket orientation. Temperature increases also cause penetration depths to be reduced more rapidly in columnar ice. For most situations encountered it is safe to assume that, since the extinction coefficient is so high the amounts of volume scattering from first year ice are effectively negligible. In contrast less saline multiyear ice enables penetration up to a metre or more in cold, low salinity situations. Such ice which has survived at least one summer's melt season has undergone fundamental changes which cause differences in scattering characteristics. The salinity distribution has been modified by brine pocket migration, under the action of a temperature gradient, and in fact near zero values above sea level appear to be primarily produced by the flushing of nearly pure surface snow meltwater downward through the ice (Weeks and Ackley, 1982). Volumetric effects are thus possible by inhomogeneities such as bubbles incorporated within the ice.

4.4.2 The particle cloud analogy for volume scattering

To approach the problem of modelling the backscatter, firstly from a snow volume, a concept proposed by Attema and Ulaby (1978) is adapted. Snow particles are likened to a cloud of scatterers in air, each of which are identical and uniformly distributed throughout the volume: these authors model backscattering from a vegetation canopy in the same manner.

If snow does not contain liquid water the imaginary part of the relative complex dielectric constant is negligible, and the relative dielectric constant or relative permittivity is a function of snow density. At normal incidence the magnitude of the relative dielectric constant varies between 1.4 and 2.2 under normal dry snow conditions, thus giving values for the power reflexion coefficient of between -36 dB and -26 dB. The actual power reflexion coefficient at the air/snow interface is 0.02 when $\epsilon' = 1.8$, and is 0.08 for ice with $\epsilon' = 3.15$. In both cases over 90% of the incident wave is transmitted across the air/snow interface. Additionally, since the power reflexion coefficient is so low, contributions due to multiple internal reflexions between the upper and lower boundaries of the snow layer may be excluded. If we ignore multiple reflexions and multiple scattering within the snow volume then the volume scattering coefficient $\sigma_v$ is given by:

$$\sigma_v = N \sigma_b$$  \hspace{1cm} (4.20)

where $N \text{ (m}^{-3}\text{)}$ is the number of scattering particles per unit volume and $\sigma_b \text{ (m}^2\text{)}$ is the scattering cross-section of an individual scatterer or snow particle. It should be noted that $\sigma_v$ has the dimension of reciprocal length (m$^{-1}$), but it represents physically the backscattering

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A snowpack modelled as a cloud of randomly distributed scatterers. The angle of incidence is $\theta$, the angle of refraction into the medium is $\theta'$, the reflexion coefficient at the surface is $R_{as}$, $V$ is the volume contributing to scatter, $\delta_p$ is the penetration depth, and $k_e$ is the extinction coefficient.

cross-section ($m^2$) per unit volume ($m^{-3}$). Also the volume scattering coefficient is often called the radar reflectivity.

If the particle size is much smaller than the wavelength of the incident wave then the Mie expression for the scattering efficiency (or normalised radar backscattering cross-section) of a spherical particle of known radius can be reduced to a more simple expression known as the 'Rayleigh approximation' (van de Hulst, 1957; Ulaby et al., 1982). Making the assumption that individual scatterers are spherical enables calculation of the backscatter cross-section for an ice particle. From Kerr (1951) it can be shown that

$$\sigma_b = 4\pi r^2 \chi^4 |K|^2 \tag{4.21}$$

where $r$ is the particle radius, and $\chi$ the dimensionless Mie size parameter is given by $\chi = 2\pi r / \lambda$, and thus;

$$\sigma_b = 4\pi r^2 \left( \frac{16\pi^4 r^4}{\lambda_0^4} \right) |K|^2$$

$$= \frac{64\pi^5 r^6}{\lambda_0^6} |K|^2 \tag{4.22}$$

where $K$ is a complex quantity defined in terms of the complex dielectric constant of a particle relative to the background medium

$$|K|^2 = \left[ \frac{(\epsilon^* - 1)}{(\epsilon^* + 2)} \right]^2 \tag{4.23}$$
For an ice particle suspended in air the condition \( \varepsilon''/\varepsilon' < 0.1 \) is sufficient that the dielectric loss \( \varepsilon'' \) can be neglected, thus \( \varepsilon^* \approx \varepsilon' \approx 3.15 \) and the value of \( | K |^2 \) is approximately 0.174.

Figure 4.8 illustrates the situation where an incident ray is transmitted through the snow surface and penetrates a known depth into the snow volume \( V \). The angle at which the ray is refracted into the snow medium \( \theta' \) is determined by Snell's Law. If we apply the theory of Attema and Ulaby (1978) to a dry snow volume then the following expression gives the backscattering coefficient from the snow volume alone;

\[
\sigma^o_v (\theta') = \frac{\sigma_w \cos \theta'}{2k_e} \left[ 1 - \exp(2k_e d \sec \theta') \right]
\]  

where \( k_e \) is the extinction coefficient (nepers m\(^{-1}\)) in the snow volume, and \( d \) is the depth at which scattering takes place.

By Snell's Law;

\[
\sin \theta' = \frac{\sin \theta}{n} \tag{4.25}
\]

but the refractive index \( n \) is related to the dielectrics by \( n = \sqrt{\varepsilon^*} \). Since the relative dielectric losses are negligible for dry snow \( \varepsilon' \) may be substituted for \( \varepsilon^* \), therefore

\[
\sin \theta' = \frac{\sin \theta}{\sqrt{\varepsilon'}} \tag{4.26}
\]

where \( \theta \) is the angle of incidence.

According to Ulaby et al. (1982), and Ulaby et al. (1984), if scattering losses are ignored then \( k_e = 1/\delta_p \), where \( \delta_p \) is given by equation 4.19 for dry snow. Inserting typical values for a snow volume of density 500 kg m\(^{-3}\), the extinction coefficient is normally of the order of 1. Equation 4.24 may be expressed in terms of the one-way loss factor \( (L(\theta')) \) of the snow volume

\[
L(\theta') = \exp(k_e d \sec \theta') \tag{4.27}
\]

thus enabling it to be simplified to

\[
\sigma^o_v = \frac{\sigma_w \cos \theta'}{2k_e} \left( 1 - \frac{1}{L^2(\theta')} \right) \tag{4.28}
\]

If for simplicity the snow depth is initially assumed greater than the penetration depth (ie. by setting \( d = \infty \)) then the contribution of surface scattering from any interface beneath the snow layer is zero. Ulaby et al. (1984) derive a final equation which may be used for calculating the coefficient of backscatter from a dry snowpack volume;

\[
\sigma^o_{pack}(\theta') = T^2_{sa}(\theta) [\sigma^o_v(\theta') \]  

\[
\tag{4.29}
\]

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Figure 4.9  Variation in the volume backscatter coefficient $\sigma_v^o$ for a dry snow layer of infinite depth. Curves are illustrated for snow with the values 1.8 and 0.006 for $\epsilon'$ and $\epsilon''$, and snow particle radii of 0.5, 1.0, and 1.5 mm.

where $T_{sa}$ is the transmission coefficient at the air/snow interface, and $T_{sa}^2(\theta) = (1 - R)^2$. When the depth may be considered infinite (ie. for depths in excess of $\delta_p$) the loss-factor $L$ has a large magnitude, enabling equation 4.28 to be simplified to

$$\sigma_v^o = \frac{\sigma_v \cos \theta'}{2k_e} \quad \{4.30\}$$

Figure 4.9 illustrates the variation in the volume scatter coefficient for an infinitely deep snow layer with a relative dielectric constant of 1.8, which is typical of a snow density of 500 kg m$^{-3}$. Snow particles are assigned three different radii with the assumption that the snow is packed such that all adjacent particles touch at a point, then the number of scatterers per unit volume $N$ in equation 4.20 is simply $(1/2\pi)^3$. It is evident that the resulting curves are dominated by a combination of the cosine of the angle of incidence and a function of $r^3$, for a medium of given dielectric composition. These theoretical results show that the larger the scatterers, in this case the snow grains themselves, the larger the component of $\sigma_v^o$ at all incidence angles.
4.5 Terrain backscattering model

Backscattering from snow covered terrestrial terrain realistically consists of contributions by surface scattering at the air/snow interface, volume scattering from the snow layer, and surface scattering by internal layers - or at the snow/ice interface (as in the case of sea ice with a shallow snow layer).

Although elaborate theoretical models have been developed to characterise the backscattering behaviour of snow covered terrain (Fung and Eom, 1981; Shin and Kong, 1981), in practice these models are difficult to apply when detailed knowledge is not available on the statistical properties of the upper and lower boundaries of the snow layer and of the particle size distribution of the scatterers (ie. ice crystals in the case of dry snow). In this section the relative contributions of surface and volumetric backscatter are evaluated for a variety of surfaces encountered by an altimeter. Using the dielectric data in the previous chapter enables a reasonably comprehensive description of terrain scattering properties in the context of polar surface media to be given.

Section 4.3.3 discusses a model for surface scatter which is applicable in a composite ‘three layer model’ for predicting scattering characteristics. To this the volume scatter model is added for completion. In a simplified three layer model there are distinct boundaries at the upper surface, at internal layers within the snow volume, and at a lower surface if one exists. These interfaces necessitate that backscatter is taken into account from each boundary. Power is returned to the altimeter by scattering at these boundaries rather than by a pure reflection process, but also consists of a component of volume scatter described in the previous section. The following equation expresses the total coefficient of backscatter of such a three-layer model as a combination of backscatter from each source.

\[
\sigma^{o}_{\text{total}}(\theta) = \sigma^{o}_{sa}(\theta) + T^{2}_{sa}(\theta')\left[\sigma^{o}_{v}(\theta') + \frac{\sigma^{o}_{si}(\theta')}{L^2(\theta')}\right] \tag{4.31}\]

\(\sigma^{o}_{sa}(\theta)\) is the contribution from the surface scattering at the snow/air interface, and may be calculated using the surface scattering model discussed in section 4.3.3. \(\sigma^{o}_{v}(\theta')\) is due to volume backscattering from the snow layer, given a power transmission coefficient at the snow surface \(T_{sa}\); where \(T_{sa}(\theta') = (1 - R)^2\). \(L(\theta')\) is the one-way loss factor of the snow layer (see equation 4.27), and \(\theta'\) is the transmission angle in the snow layer. Finally \(\sigma^{o}_{si}\) is added for completion to account for surface scatter from an ice layer of known dielectric properties, beneath or within the the snow layer. It becomes effective under the condition where \(d < \delta_p\).

Figures 4.10 a,b,c,d,e, and f illustrate the results of the three-layer model for differing surface roughnesses, dielectric properties, and layer depths. The values for rms slope \(\beta = 0.05\) rad, \(\beta = 0.15\) rad, and \(\beta = 0.45\) rad correspond to surfaces which are smooth, medium
rough, and very rough. Figure 4.10a represents a snowpack of density 500 kg m\(^{-3}\) of uniform
dielectric properties without any internal horizons, and thus effectively infinite depth. Snow
grain diameter is 1.0 mm and the half-space or volume backscatter contribution is as a result
-17 dB. The three curves in this and following plots show the scatter signature for each
surface roughness. Experimental results from Moore et al. (1980) for similar snow dielectric
properties are plotted alongside the model curves in Figure 4.10a. They illustrate that the
model reproduces the scattering characteristics of a snow layer reasonably faithfully, and
although the results of Moore et al. (1980) indicate a slightly higher half-space contribution,
this could be achieved by increasing the snow grain size.

Figure 4.10b shows scattering from a snowpack of limited depth \((d = 1 \text{ m})\), of similar
dielectric properties as that in 4.10a. Beneath the snow is a smooth layer of glacial ice with
relative permittivity 3.14. Curves show a small increase at all grades of snow surface roughness
between 0° and 10° incidence as a consequence of reflection from the surface of the ice (where
the reflection coefficient \(R(0) = 0.019\)). This indicates that ice layers, inhomogeneities, or
dielectric interfaces within the snowpack may cause a large component of backscatter providing
that the snow layer is not too deep to mask these effects.

Figure 4.10c represents the situation for a dry late-spring snowpack with an ice layer
at 1 m depth. Destructive metamorphism is accounted for with a snow grain diameter of
2 mm, double the size in both previous examples. The dielectric values represent a snowpack
of density 500 - 600 kg m\(^{-3}\). The higher relative permittivity \(\epsilon'\) and larger grain size cause
the volume backscattering component to increase by 12 dB (ie. a factor of 15 increase),
resulting in a half-space value of -5 dB. The higher reflection coefficient at the snow surface
\((R = 0.038)\) increases the surface scatter response by 9 dB at normal incidence. Contributions
by reflection and scattering at the internal snow/ice interface (where \(\beta_{\text{ice}} = 0.05 \text{ rad}\) are
lower than in the previous example because the dielectric mismatch between upper and lower
layers is smaller. Experimental results by Fung et al. (1980) for a normal snow surface, and
wind generated rough surface of similar dielectric properties correspond well with the model
predictions. Apart from the higher value of \(\sigma^0\) at normal incidence for the experimental rough
surface results, the range of typical values for snow surface roughness lie between an rms slope
of 0.05 rad at the smoothest, to 0.15 rad for rough wind generated microrelief.

In contrast to Figure 4.10c, Figure 4.10d shows the predicted backscatter response of
a 10 cm deep new layer of low density snow above a thin ice layer representing ice lenses,
superimposed ice upon the former surface, or a refrozen surface. Destructive metamorphism
is minimal and snow grain diameter is reduced to 0.5 mm. The snow layer, simulating a density
of 200 kg m\(^{-3}\), contributes a negligible amount of volume scatter (below -30 dB), and that
from the snow layer beneath the ice is also assumed negligible and ignored. Most dominant are
Figure 4.10  Model predictions of angular variation in $\sigma^0$ at different snow surface slopes ($\beta$) for a variety of physical and electromagnetic conditions. (a) shows results for a dry snow surface of infinite depth. (b) is for a dry snow covered ice layer; (c) is for higher density snow of similar depth upon ice; (d) is for a shallow low density snow layer upon an ice layer; (e) is for a wet snowpack; and (f) is for a bare freshwater ice surface.
the combined surface scattering components from the snow surface of varying roughness and the smooth thin ice layer (of a fixed rms slope $\beta = 0.05$ rad). Since over 99% of incident energy penetrates and extinction losses in the snow are minimal the majority of power is returned by the internal snow/ice interface. Scattering at the air/snow interface would appear to have insignificant effect upon scattering signature between 0° and 10°. Beyond an incidence angle of 10° from the normal the largest amounts of backscatter come from the rough snow surface. As expected in these cases from Figure 4.2b $\sigma^o$ decays less rapidly for the rougher surfaces.

Figure 4.10e illustrates the situation of a wet snowpack of density $500 \text{ kg m}^{-3}$ with a depth greater than the penetration depth (ie. $d = \infty$). High relative permittivity and dielectric losses are caused by a free water content near the limit of the pendular regime, and the resulting reflexion coefficient at the snow surface is 0.072. As attenuation or extinction losses within the medium are large the volume backscatter component is below -20 dB and is thus minimal. Experimental results by Fung (1981) for rough and smooth wet snow surfaces show marked similarity with the model predictions for values $\beta = 0.15$ and $\beta = 0.05$ respectively, but only at incidence angles less than 10°. The less rapid decay thereafter in the results of Fung (1981) would seem to indicate that the water content, simulated by high dielectric losses in the model, is greater than in the snowpack they investigate. This argument is supported by the observation that fractionally lower water contents reduce dielectric losses and the loss factor $\tan \delta$ by large amounts, with corresponding reductions to the extinction coefficient of the snow. The effect in the model is to raise the half-space contribution from the upper parts of the snowpack.

Finally, the backscatter response of a bare glacier ice surface is investigated, for contrast to the wet snow results. Bare ice was found to have a similar reflexion coefficient to wet snow surfaces in Chapter 3, and it is not surprising therefore to find that the variation in $\sigma^o$ with $\theta$ in Figure 4.10f is almost identical with Figure 4.10e. The response to varying degrees of roughness is similar, and the only significant difference is that glacier ice has lower dielectric losses and so contributions by scattering from bubbles and inhomogeneities within the upper layers of the ice. These cannot be represented accurately here, but nonetheless the model demonstrates a half-space component of the order of 6 dB higher than in the wet snow example.

In all cases in Figure 4.10 near-normal incidence surface scatter dominates the scattering pattern, with volume scattering merely providing a background level of some $-10$ dB or less. When there is a highly reflective layer beneath the snow, it only becomes important when losses within the snow layer are minimal, or if the snow depth is shallow. The results show marked similarity to those of Moore et al. (1980), Fung et al. (1980), Stiles and Ulaby, (1980), Stiles et al., (1981), and Fung, (1981), confirming that the model is giving results of the correct
order of magnitude. In the case of wet snow Figure 4.7a shows that the penetration depth is minimal, and of the order of a wavelength. When the relative dielectric constant and loss become large both \( k_e \) and \( L^2 \) increase dramatically, and backscatter from the snow or ice volume becomes unimportant. In such cases the surface scatter contribution dominates the scattering pattern, and the magnitude of \( \sigma_{total}^\circ \) reaches its highest values, due to the increased reflection coefficient of the surface.

The conclusion is therefore that surface scatter from ice or snow terrain dominates the backscatter coefficient. Although volume effects are evident for dry snow media, the dynamic range of the RAL altimeter is too restricted to detect values below approximately 0 dB. As the backscatter coefficients of different types of snow and ice surfaces are so variable it may be possible to detect changes in the backscatter records obtained by the altimeter over terrestrial snow covered ice surfaces. Figures 4.10 a-f indicate that dry snow may possibly be distinguished from wet snow or bare ice surfaces, on the basis of the value of \( \sigma^\circ \) at normal incidence, and the relative decay of \( \sigma^\circ \) with angle may be used to categorize surface roughness.

### 4.5.1 Scattering from sea ice

The modelling of backscatter from sea ice has been reported by Parashar et al. (1978), Fung and Eom (1982), and recently by Kim et al. (1985) and Onstott (1987). In these studies, scattering by volume inhomogeneities and irregular surface boundaries has been treated in different ways. In some cases they are treated independently, while in others they are combined in a composite fashion, as in the previous section. Here the terrain scattering model is modified to treat volume scattering from within a sea ice medium which may or may not be covered by a shallow snow layer.

#### 4.5.1.1 Modifications to the scattering model

According to Figures 4.7b and c the skin depth of first year ice at 13.8 GHz is calculated to be of the order of a few centimetres or less. In fact, for saline ice, the power reflection coefficient \( R \), is shown in Chapter 3 to tend to a value of 1 at \( 0^\circ \) C, and so negligible amounts of the incident wave penetrate bare ice surfaces. In contrast, in multiyear ice, where electromagnetic losses are less, the penetration depth may be of the order of a metre (see Figure 4.7d). In such cases the effects of air bubbles found in the top layer of multiyear ice may contribute to scattering. The terrain scattering model must therefore be modified to include the effects of volume scattering from the ice volume as well as a snow layer volume. Fung and Eom (1982) discuss scattering from bubbles within sea ice volumes at Ku-band frequencies. They assume a random collection of air bubbles embedded in a homogeneous lossy medium. Typical values for the relative dielectrics of the lossy background saline ice may be taken from Chapter 3,
while the relative permittivity of the air bubbles is assumed to have the same value as free space (i.e., 1). The volume fraction of air bubbles \(f_{air}\) can be estimated using the density of sea ice \(\rho_t\) and equation 3.27 in Chapter 3. The number of scatterers per unit volume \(N\), from Fung and Eom (1982) is

\[
N = f_{air} \left( \frac{4\pi r_s^3}{3} \right)^{-1},
\]  
{(4.32)}

where \(r_s\) is the radius of the scatterers.

It is recalled that \(K\) is a complex quantity defined in terms of the complex dielectric constant of the scatterer (subscript \(s\)) relative to the background medium (subscript \(b\)). Substituting a value of \(\epsilon^*_s/\epsilon^*_b\) for \(\epsilon^*\) in equation 4.23 enables calculation of the scattering cross-section of a bubble within saline ice. When multiplied by \(N\) this gives the scattering cross-section per unit volume. Expression 4.24 is modified to derive the volume scattering coefficient of a sea ice medium, given a particular volume fraction of air bubbles, as follows;

\[
\sigma_{ice}^o(\theta'') = \frac{\sigma_{v} \cos(\theta'')}{2k_e} [1 - \exp(2k_e d \sec \theta'')]\]
{(4.33)}

Importantly, the wave passes through a saline ice background medium, rather than air. Its wavelength \((\lambda_b)\) is now \(\lambda_b = \lambda_o/\sqrt{\epsilon_b^*}\). The angle of refraction within the sea ice medium \(\theta''\) may be calculated using equation 4.25 where \(\epsilon^*\) is the average complex dielectric constant of the sea ice. Van Beek (1967) and Fung and Eom (1982) derive a formula to find the average dielectric constant \(\epsilon^*_{mix}\) of spheroids (in this case air bubbles) with complex dielectric constant \(\epsilon^*_s\) embedded within a homogeneous saline ice medium of complex dielectric constant \(\epsilon^*_b\), as follows;

\[
2(\epsilon^*_{mix})^2 - \epsilon^*_{mix} (2\epsilon^*_b - \epsilon^*_s + 3f_{air}\epsilon^*_s - 3f_{air}\epsilon^*_b) - \epsilon^*_b\epsilon^*_s = 0
\]
{(4.34)}

The quadratic may be solved for \(\epsilon^*_b\) when a value of 1 is substituted for \(\epsilon^*_s\) for air and the volume fraction of air \(f_{air}\) and the dielectric constant of the ice are known.

Ultimately the amount of backscatter from the sea ice transmitted back into the snow layer via the sea ice surface is regulated by the power transmission coefficient at the snow ice interface \(T_{si}\). The amount may be calculated by the following equation;

\[
\sigma_{ice}^o(\theta'') = T_{si}^2(\theta'')(\sigma_{ice}^o)
\]
{(4.35)}
4.5.1.2 Model results

A parametric study is undertaken using the modified three layer model with input data from sea ice and snow measurements by Tucker et al. (1987) during the MIZEX '84 experiment. These records enable a reasonably accurate description of the snow and ice dielectric properties, and thus predictions of the backscatter signatures of different Greenland Sea marginal ice types. Comparisons may then be made with RAL altimeter derived backscatter data in the following chapters.

Figures 4.11 a-f illustrate some model results for different types of sea ice mediums with variable snow cover characteristics. The half-space or volume scattered proportion of the total backscatter coefficient, as well as the ice surface scattered proportion of the coefficient are represented as dotted lines. This emphasises the relative proportions of scattering from the surfaces and snow and ice volumes. Three curves are plotted representing the combined scattering response using the same values for snow surface rms slope as in the previous section: \( \beta = 0.05 \) rad is smooth, \( \beta = 0.15 \) rad is medium rough, and \( \beta = 0.45 \) rad is very rough.

Figure 4.11a represents the typical situation found by Tucker et al. (1987) for first year floes of density 850 kg m\(^{-3}\) covered in a thin layer of snow and at a temperature below zero. A salinity value of 4\(^\circ\)/o is used (see Figure 3.10), the snow depth in the model is set to 0.08 m in this example and the ice surface is given an arbitrary value for rms slope of \( \beta_{\text{ice}} = 0.05 \) to indicate a relatively smooth snow/ice interface. The remaining parameters are indicated in Figure 4.11a. Predictions of \( \sigma^0 \) with incidence angle show that in the region 0–10° scattering is dominated by the ice surface and not the uppermost snow surface, simply because the dielectric mismatch between the saline ice and dry snow is greatest at this interface. The dry snow layer has negligible influence until beyond this 10° limit whereupon the roughest snow surfaces cause the highest magnitudes of \( \sigma^0 \). Volume scatter in the snow and ice layers together account for -19 dB. Figure 4.11b shows the situation when the first year ice parameters for Figure 4.11a are used for an ice surface of increased roughness where \( \beta_{\text{ice}} = 0.1 \) rad. The scattering response is marked with \( \sigma^0 \) reduced in each case at normal incidence by between 5 and 8 dB, and a less rapid decay in \( \sigma^0 \) between 0° and 20°. This indicates that ice surface roughness causes a critical response in the scattering model for high salinity ice types and shallow snowcovers. Figure 4.11c shows the scattering response to wetting the snow cover upon the ice surface. This situation simulates conditions of high air temperatures and large volume fractions of free water within the snow from snow surface melt. In contrast to the two previous examples the snow surface roughness now has the most dominant effect upon the variations in \( \sigma^0 \) with rms slope. The ice surface scatter is reduced in significance receding to a level below the half-space component of -16 dB.

Losses in the wet snow layer almost mask out all surface scatter from the snow/ice inter-
Figure 4.11 Modified terrain model predictions for the backscattering response of first year and multiyear ice types using different surface layers with distinct dielectric properties. Air/snow and snow/ice interface roughnesses are not shown in the diagrams but are discussed in the text. (a) first year ice with a low density dry snow layer 0.08 m deep. (b) represents the same snow characteristics with an increased ice surface roughness of 0.1 rad. (c) shows the consequences for scattering when the shallow snow layer upon a first year floe is wetted. (d) is for a bare firstyear ice surface which has experienced warmer temperatures and undergone recrystallisation at the upper surface. (e) is for a thicker multiyear ice floe with a deep dry snow layer, and (f) is the same multiyear floe when the snow layer is wetted.
face. Furthermore, for smooth wet surfaces the backscattering signature is indistinguishable from the situation in Figure 4.11a between 0 and 10°.

In Figure 4.11d the model is used to predict the variation in $\sigma^0$ for a bare first year ice surface which has undergone recrystallisation in the upper 0.03 m. The top layer in the model is assigned dielectric values typical of pure ice so as to simulate the conditions for a floe upon which the snowcover has melted and refrozen. The curves for differing ice surface roughness show that there is a large component of backscatter from the pure ice surface, yet the largest component of $\sigma^0$ occurs by penetration of the pure ice and scattering at the saline ice surface beneath. Different values of rms slope cause large variations in the rate of decay of $\sigma^0$ with $\theta$, and the value of the backscatter coefficient at normal incidence falls from a maximum of 12 dB or more for smooth surfaces to less then 2 dB for very rough surfaces. Volume scattering is again minimal at -14 dB.

Multiyear ice observed by Tucker et al. (1987) in the marginal ice zone of the Greenland Sea is different from first year ice in a number of ways. Figure 3.10 shows that salinities are lower by about 3 ppt, resulting in a lower dielectric constant. In general ice measurements indicated a mean thickness of 3.2 m and a density of around 700 kg m$^{-3}$, and so similar values are used in the model. Additionally, as larger penetration depths are expected in lower salinity ice, bubbles take effect as Rayleigh scattering centres (Onstott et al., 1987): bubbles are assigned a radius of 1.2 mm in the model upon observations that this was typical of the order of size found in Fram Strait multiyear ice. Tucker et al. also report that snowdepth averaged a value of 28.5 cm upon multiyear floes. Figure 4.11e shows the scattering signatures for multiyear ice with a dry snow cover and a fixed ice surface rms slope of $\beta = 0.15$ rad. Incident energy is observed to penetrate the snowcover and the ice surface is shown to contribute a value of -4 dB to $\sigma^0$. The peak value at normal incidence for a smooth snow surface layer is less than 4 dB, and the more likely medium rough and rough situations result in a gradual decay of $\sigma^0$ with incidence angle (less than 0.1 dB/Degree): these results show a marked contrast to those of first year ice examples in Figures 4.11a and b. Volume scatter is observed to play an important role and the combined snow and ice half-space contributions set a background level of -7 dB. Finally, Figure 4.11f illustrate the model predictions for when the snow layer upon the multiyear ice surface is wetted with pure water. This confirms the observations made in Figure 4.11c that the snow layer masks the signature of the ice type completely. Since in this case snow depth is greater, no appreciable ice surface scatter is recorded. As a result the snow surface scatter dominates and the pattern is identical between 0 and 10° to that in Figure 4.11c.

Measurements undertaken by Burns et al. (1986) indicate that when sea ice measurements were made with the RAL altimeter, snow on the surfaces of ice floes had grain sizes
between 1 and 5 mm. Thus the model value of 2 mm diameter is cor-
findings. Snow density had a mean value of 500 kg m\(^{-3}\) throw-
sensing observations and dielectric values used as model para-
possible to ones calculated at this density in Chapter 3. Fur-
fractions fell in the 0 to 4\% range, much of the variation in m. 
for by spatial variations in snow conditions. On the day of obs.
prior to the altimeter observations, air temperatures were below zero (\(\text{\textdegree} C\)) 
all but one day. On the day of the flight over sea ice the air temperature 
therefore likely that the amounts of free water within the snow surface layers 
were minimal. It is suggested therefore that on the day of altimeter measurements the 
scattering signatures correspond most closely to the dry snow model predictions in Figur. 
4.11a, b, and e, although it is recognised that variations in the spatial physical properties 
will ultimately modify these to differing degrees. There is some possibility that the snow may 
have an insulating effect, for instance, which the model is not presently designed to cope with.

On this note a caveat should be added for the use of the model. Kim et al. (1984) calculate 
for a 1m thick first year ice floe with 10 cm snow cover that an air temperature of \(-30^\circ\text{C}\) 
would be reduced to \(-17.5^\circ\text{C}\) by insulation (using equations by Nakawo and Sinha (1981) for 
an ice surface beneath a snow cover). The consequence is that the dielectric constant of the 
snow covered ice surface is increased to 3.29 from 3.21, when the ice has a salinity of 10 ppt.

At normal incidence the resultant Fresnel power reflection coefficient changes from 0.084 to 
0.080, and the consequences for scattering are not small enough to be neglected. The model 
cannot be used therefore, without a rigorous investigation of the dielectrics of the layers prior 
to its implementation. Onstott et al. (1982) measure the effects of variable snow depth upon 
scattering signatures of a variety of ice types. As the thickness of the snow layer upon thick 
first year ice increases, the absolute level of the radar cross-section also increases. Returns 
enhanced in such a manner had increases ranging between 1 and 4 dB in most cases with an 
 extreme example of 8 dB.

Although there are a number of shortcomings recognised in this approach, scattering 
measurements made by Onstott et al. (1979), Kim et al. (1984), and most recently Onstott et 
al. (1987) actually during MIZEX '84, correspond well with simulated coefficients using the 
modified equation 4.31, for equivalent snow and ice parameters. Values for \(\sigma^\circ\) at angles close to 
normal incidence are generally between 0 and 10 dB for both first year and multiyear ice types, 
yet first year ice generally shows higher scattering coefficients at these angles. Nevertheless, 
as expected the angular decrease in backscatter coefficient is much more rapid for first year 
ice. Notably, near-normal incidence scattering is dominated by the presence of the rough 
surface scatter from either the snow surface or the bare ice surface, and is governed by the 
degree of roughness. It is only beyond angles of incidence of 25\^\circ that volume scatter from
between 1 and 5 mm. Thus the model value of 2 mm diameter is consistent with experimental findings. Snow density had a mean value of 500 kg m\(^{-3}\) throughout the period of remote sensing observations and dielectric values used as model parameters correspond as closely as possible to ones calculated at this density in Chapter 3. Furthermore, most snow water volume fractions fell in the 0 to 4\% range, much of the variation in measured values being accounted for by spatial variations in snow conditions. On the day of observations and for the week prior to the altimeter observations, air temperatures were below zero (Burns et al., 1986) on all but one day. On the day of the flight over sea ice the air temperature was \(-3^\circ\) C. It is therefore likely that the amounts of free water within the snow surface layers on that day were minimal. It is suggested therefore that on the day of altimeter measurements the sea ice scattering signatures correspond most closely to the dry snow model predictions in Figures 4.11a, b, and e, although it is recognised that variations in the spatial physical properties will ultimately modify these to differing degrees. There is some possibility that the snow may have an insulating effect, for instance, which the model is not presently designed to cope with. On this note a caveat should be added for the use of the model. Kim et al. (1984) calculate for a 1m thick first year ice floe with 10 cm snow cover that an air temperature of \(-30^\circ\) C would be reduced to \(-17.5^\circ\) C by insulation (using equations by Nakawo and Sinha (1981) for an ice surface beneath a snow cover). The consequence is that the dielectric constant of the snow covered ice surface is increased to 3.29 from 3.21, when the ice has a salinity of 10 ppt. At normal incidence the resultant Fresnel power reflexion coefficient changes from 0.084 to 0.080, and the consequences for scattering are not small enough to be neglected. The model cannot be used therefore, without a rigorous investigation of the dielectrics of the layers prior to its implementation. Onstott et al. (1982) measure the effects of variable snow depth upon scattering signatures of a variety of ice types. As the thickness of the snow layer upon thick first year ice increases, the absolute level of the radar cross-section also increases. Returns enhanced in such a manner had increases ranging between 1 and 4 dB in most cases with an extreme example of 8 dB.

Although there are a number of shortcomings recognised in this approach, scattering measurements made by Onstott et al. (1979), Kim et al. (1984), and most recently Onstott et al. (1987) actually during MIZEX ‘84, correspond well with simulated coefficients using the modified equation 4.31, for equivalent snow and ice parameters. Values for \(\sigma^o\) at angles close to normal incidence are generally between 0 and 10 dB for both first year and multiyear ice types, yet first year ice generally shows higher scattering coefficients at these angles. Nevertheless, as expected the angular decrease in backscatter coefficient is much more rapid for first year ice. Notably, near-normal incidence scattering is dominated by the presence of the rough surface scatter from either the snow surface or the bare ice surface, and is governed by the degree of roughness. It is only beyond angles of incidence of 25\(^\circ\) that volume scatter from
the sea ice dominates scattering, and then its angular trend is controlled by both the rough surface boundary and the average dielectrics of the saline ice with bubbles. Onstott et al. (1982) confirm with radar cross-section measurements at Ku-band that although backscatter from sea ice is not purely surface scatter, volumetric scattering effects are only observed as extremely low signal levels. These effects are below the dynamic range of most instruments, and so as a result, what is observed with the RAL altimeter is affected most significantly by the surface scattering characteristics of both snow and saline ice.

The model provides useful estimates for both snow and sea ice scattering providing that realistic and consistent parameters are input. At angles close to nadir and normal incidence, its results suggest that it is possible to distinguish between first year and multiyear ice on the basis of backscatter behaviour. Between 0 and 10° RAL altimeter calibrated backscatter measurements can be used to give the magnitude of σ° at normal incidence, and to estimate the variation in σ° with θ. Using this technique, first year ice and multiyear ice may be identified. Unfortunately, to date, there are very few published results from MIZEX '84 which may be used to support the modelling results discussed here. It is hoped that the validity of these modelled signatures will be confirmed and tested in future analyses of MIZEX data.
CHAPTER 5

ASPECTS OF RADAR ALTIMETRY IN MARGINAL SEA ICE AREAS

5.1 Introduction

This chapter and the next investigate aspects of radar altimetry over two distinct types of polar ice surfaces in the Arctic. Principal or characteristic features of altimeter pulse waveforms are dealt with and interpretations and explanations are made in the context of glaciologic phenomena.

Chapter 5 is an investigation of radar altimetry over sea ice surfaces, and is based on data collected during MIZEX '84 by the RAL altimeter. Although altimetric data were collected over sea ice areas several times during the MIZEX campaign, the coincidence of cloud-free photography and a number of sources of aircraft data on the 30th June 1984 has led to its selection by most scientists involved with MIZEX '84 remote sensing investigations (McIntyre et al., 1986). It is for this reason, and the fact that instrument malfunctions were minimal, that results presented in this chapter are from the flight on that day (ie. flight 12).

A series of questions remain unanswered about sea ice and its influence on sensible heat exchange in the oceans, and between ocean and atmosphere. Aperiodic year to year variations in the position of the ice boundary reflect feedback in the mass balance of the sea ice cover. Remote sensing instruments, particularly radars, have an important role to play in deriving geophysical parameters of use in modelling studies, and which further an understanding of sea ice conditions and dynamics. The Marginal Ice Zone (MIZ) represents one of the most active regions within the global sea ice cover, and impinges on, and provides ice to zones of economic activity for the offshore engineering and fishery industries. Data from the RAL airborne altimeter, and forthcoming satellite altimeters, will enable users and scientists to propose, define, and develop operational satellite data products. MIZEX is unique in that it is yielding the first multi-sensor comparisons from remote sensing instruments in important sea ice areas. Combination of microwave data, airborne photography, and some surface information, will substantially improve our knowledge of the type of microwave response that occurs, and what factors govern that response.

This chapter discusses several main themes using RAL airborne altimeric data: (1) how a radar altimeter responds over sea ice/water mixtures; (2) scattering phenomena common to distinct regions within the MIZ; (3) the discrimination of ice types on the basis of model
results in chapters 3 and 4; (4) the extraction of useful geophysical parameters from altimeter waveforms; and (5) algorithms capable of deriving sea ice products from future satellite data.

5.2 Flightplan: 30th June 1984

The flightplan in Figure 5.1 consists of five 17-minute legs running east/west, starting in the north at 79°29.8′N 5°57.1′E, and following a raster pattern, each limb of which is separated by 10 nautical miles (Gloersen et al., 1984). This permitted intersection with the predicted position of the Polarstern, given the previous day at 2400Z. At the end of the fifth leg, the plan was to fly on a bearing of 315° for a transect into heavy multiyear ice, but that had to be aborted in favour of extending legs one and two to 10°50′W in order to fly over Polarstern as she was struggling to make headway through heavy ice at 79°24′N 10°42′W.

Both the RAL altimeter and the cameras were operational throughout the flight. The weather was also exceptionally clear, permitting excellent aerial photographic results at a scale of 1:120,000, to be used in conjunction with the analysis of the altimetric data.

Figure 5.1 CV-990 flightline during flight 12 on June 30, 1984 over the East Greenland Sea marginal ice zone. Time ticks are approximately every 10 minutes. The overlapping CV-580 Synthetic Aperture Radar image mosaic in Plate 5.1 is delineated.
5.3 Background

The properties and composition of sea ice in the East Greenland Sea are of paramount importance when interpreting altimetry from this area. Transmitted pulses are reshaped by scattering from both the ocean surface between floes, and the surfaces and volumes of ice floes themselves.

Appendix A describes and illustrates the formation and typical characteristics of sea ice forms found in this region. Theoretical results presented in Chapters 3 and 4 on the reflective and scattering properties of sea ice give insight into processes leading to altimeter responses in the MIZ. Notwithstanding these results, the extraction of useful parameters from altimetry over sea ice is difficult. The main problem lies in the variability of the nature of the waveforms and the complexity of the nature of the surfaces being sensed. A considerable lack of experimental data are available with which to characterise the responses from different individual surfaces at near-normal incidence (McIntyre et al., 1986). It is clear that further experimental work needs to be done in Ku-band, on near-normal incidence scattering from composite ice/water surfaces, to make more accurate interpretations of altimeter data than those presented in this chapter.

5.3.1 East Greenland sea ice conditions during MIZEX-84

The East Greenland MIZ exhibits varying degrees of compactness, and the ice edge displays changes in definition, both daily and seasonally. Seldom does it exist as a straight north-south line, but follows a tortuous course with bays and projections extending miles into and out of the pack. Gyres and a combination of wind and oceanographic factors force large eddies which spiral their way southwards under the predominant flow of the East Greenland Current. Ice is observed to move several km in a day under the influence of this strong current.

The synthetic aperture radar (SAR) image mosaic in Plate 5.1 illustrates the overall ice conditions on 30th June for the area delineated in Figure 5.1 1. The ice edge, where overflown by the altimeter, is outside the SAR window, and was ill-defined and located at 2° W. A number of complicated eddy structures were present, the largest of which was located at 78° 45' N 2° 30' W, just south of the main flight pattern. This was avoided on all but the final flight leg which crossed the western edge of the eddy.

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1 The image, processed by ERIM, is formed of 7 individual swaths acquired within a 3 hour period immediately after the altimeter flight.
Plate 5.1  Synthetic aperture radar image mosaic acquired on 30th June 1984 by the Canadian Convair 580, by courtesy of the Environmental Research Institute of Michigan (ERIM).
The MIZ was composed of three distinct regions: an inner zone of large multiyear and first year floes, a transition zone of uniformly broken smaller floes, and a complex region of brash and small floes near the extreme ice edge. Observations of multiyear ice from Polarqueen in the inner region, indicated floes which were 2-5 m thick and typically a few hundred metres or more across (Maykut, 1984). First year ice floes, in contrast, were much smaller, tending to get broken up between heavier floes, with the resulting brash ice choking leads. The transition zone (approx 5-20 km in width) is characterised by fairly uniform floe sizes which decrease with proximity to the ice edge, in response to wave propagation from the open ocean. Ice concentration is relatively high (70-90%), showing only a slight tendency to decrease with average floe size. Normally, in the transition and internal zones floe geometry is dominated by interfloe collision and pressure. While penetrating long period swell may lead to cleanly fractured edges, such features are less common than the more rounded edges created by the former process. Shipborne observations from Polarqueen indicated that leads between floes were usually free of brash (Maykut, 1984). Instead of crushing between floes, the primary cause is presumed to be the effects of wave erosion and melting of small pieces of ice (due to solar energy absorbed by the water). On some days the ice edge was very distinct, being composed of a narrow (1-2 km) region of nearly 100% concentration, while on others it was very diffuse and contained bands, streamers, and fingers or detached patches of small floes. This variability was caused by changes in the prevailing wind direction and strength.

Ice conditions logged after observations from Polarstern suggest the pack was inhomogeneous in terms of age. It consisted of a mixture of first year, multiyear ice, and composite ice floes, though multiyear ice was most abundant (Tucker et al., 1984; 1987). Much of the first year ice is deformed before entering Fram Strait, as indicated by many ridges (in Plate 2a) and rubble piles of first year ice located along the edges of multiyear floes. Composite floes consisted of multiyear ice either embedded in first year ice or multiyear floes to which remnants of first year ice were still attached. First year ice thicknesses ranged from 0.38 m in a refrozen lead, to a maximum freeboard of 2.36 m. Multiyear, in contrast, ranged from 1.74 m to 5.74 m, whilst thicknesses greater than 3.50 m are attributable to old pressure ridging, with sail heights in general varying between 3 and 5 m. The mean salinity of first year ice floes was 4 °/oo, whilst for multiyear ice it averaged 2.1 °/oo (see Figure 3.10 from Tucker et al., 1984; 1987). Low salinities and a crystal texture indicative of fresh ice were observed in the upper 10 cm of several first year ice floes implying that some melting and refreezing had occurred prior to coring observations. At ground level, it was difficult to distinguish between first year and multiyear ice floes visually, because of the snow cover masking ablation features. The average snow depth on multiyear ice floes was 0.28 m, whilst on level first year floes it averaged only 0.08 m (and never exceeded 0.20 m).
Plate 2a  Typical first year ice floe surfaces in the ice edge zone. The floe in the centre of the frame is approximately 20 metres in diameter.

Plate 2b  Greenland Sea interior MIZ ice conditions during MIZEX ‘84. The maximum height of the pressure ridge in the foreground is approximately 2.5 metres.
Generally, first and multiyear ice floes could be distinguished on the basis of snow depth, and from salinity of the uppermost ice. Plate 2b illustrates the surface characteristics exhibited by typical first year ice. Floes in the near ice edge region, which undergo continual buffeting, often have highly saline surfaces (over 6 °/oo) from saltwater spray, and rims of rubble around many of their edges. These rims are of varying height, but may be as high as 0.5 m when blocks are rafted onto a floe edge. Otherwise the topography of floes is smooth with clean broken edges due to wave fracture, in contrast to the rounding occurring due to interfloe collisions. In summary a wide variety of edge geometry and surface characteristics may be found in this region.

Melting and refreezing had occurred during warm spells before the 30th June, although air temperatures on that day were several degrees below zero (Burns et al., 1986). Surface observations by Maykut (1984) showed that the first year floes with reduced snow depths had suffered rapid ablation. The result was that many first year floes had patchy snowcover indicating melting of the shallow snow layer. It is evident from the aircraft images that on many of these floes the snow cover may be as little as 50%, with meltponds occurring in places (whether frozen or otherwise). Maykut (1984) and Grenfell (1984) observe that melting and refreezing caused the surface layers of the bare ice to be formed of lower salinity superimposed ice. This would account for multiyear microwave signatures recorded by Gloersen (1984) over most of the MIZ, and multiyear-like microwave signatures over certain areas of first year ice. From the aircraft photography it is evident that floes in the north and northeast of the area are generally large (over 1km diameter), with an even snowcover and extensive pressure ridging. Fedor (1984) observes that 90% of the ice cover was composed of multiyear ice. The further south one gets the more evidence of surface ablation occurs. Rotten floes with patchy snowcover are more frequently seen, having a large degree of surface roughness (Ross and Tomchay, 1984).

5.3.2 Principles of altimeter operation over sea ice

The schematic diagram in Figure 5.2 illustrates the interaction of an individual pulse shell with a relatively smooth planar surface typified by the MIZ. From the point in time at which the leading edge of the pulse strikes the surface it moves outwards radially. The trailing edge follows fractionally later resulting in the illuminated footprint consisting of first a circle of growing area and then an annulus of increasing radius and decreasing width (but constant area).

If the sea surface slope is assumed negligible, and the surface roughness (in the form of waves and floe surface ridging and hummocking) is no more than a few metres, then the surface may be represented by a horizontal plane surface with minor perturbations superimposed.
Secondly, backscattered power is assumed proportional to the area of surface illuminated at any instant. Given a surface with metre-scale roughness of the types described, surface slopes will normally be randomly distributed. If, as discussed in Chapters 3 and 4, the backscattered power originates from Fresnel reflexion from smooth randomly distributed surface components, then each surface facet contributing to received power must be oriented normally to incident energy. Thus, as the annulus progresses outwards in Figure 5.2, successively larger surface slopes are required for scattering to take place. And, since more specular facets are likely to be oriented near-horizontally, the corresponding backscatter will fall rapidly from a maximum at nadir and normal incidence, with increasing incidence angles off-nadir (corresponding with time t₀ to tₙ in Figure 5.2).

![Figure 5.2](image-url)

**Figure 5.2**  Schematic diagram illustrating interaction of an altimeter pulse with a horizontal planar sea ice surface. The upper section shows the increasing incidence angle \( \theta_{1-5} \), and the corresponding time delays relative to receipt of peak power at \( t_0 \). The lower section shows the footprint areas contributing to a particular range bin in each recorded waveform, and the change in backscatter coefficient as a function of delay time \( t \) (and/or incidence angle).
Digitally recorded waveforms stored as power values in incremental time delay gates (or range bins) contain a record of the variation in backscatter with incidence angle. Using aircraft pitch and roll ADDAS records, data are corrected for antenna pointing errors, and given the antenna response pattern, backscatter records are corrected to eradicate effects of antenna beam-attenuation. From these records, diagrams of the angular (polar) scattering function may be plotted, and inferences made regarding characteristics of backscatter over the range of incidence angles. These, in turn, are closely related to the surface dielectric and roughness properties and the distribution of backscattering slope facets.

NB. not to scale.

![Diagram](image)

**Figure 5.3** The error $\delta \theta$ in calculating the variation in $\sigma^0$ with incidence angle $\theta$ is caused by pressure ridging or large-scale surface roughness of height $h_r$ above the mean surface. For a ridge displaced from nadir a differential path length $\Delta R$ is required for it to appear as if at nadir, where $\Delta R = h_r / \cos \delta \theta$. For an aircraft height $H$ the angular error $\delta \theta$ is simply $\cos \delta \theta = (H - h_r) / H$, and a typical ridge height of 3 m yields an error of 1.4°.

Assumptions may break down over sea ice in two cases, in areas where; (a) a large proportion of backscatter originates from volumetric scattering sources, or (b) when surface roughness is large enough for an off-nadir ridge, for example, to be nearer than the nadir point on the mean surface. In the former case the derived angular backscattering function must be treated with care, while in the second case the situation is illustrated in Figure 5.3. When pressure ridges are large, it may be that the first returned echo does not originate from normal incidence. Figure 5.3 shows the maximum error $\delta \theta$ in the calculation of a dB vs $\theta$ polar scattering diagram: for a ridge height of 3 m the maximum possible error is 1.4°. This uncertainty error may exist over all specular surfaces with large-scale perturbations superimposed upon it, and is the same for a sea surface with a maximum waveheight of 3 m.
Thus for the typical range of ridge heights observed in the MIZ on June 30th, the errors are less than 1.5°.

For any nadir-viewing sensor the strongest returns will occur when the surface is a flat specular reflector, as over calm seawater. Roughness causes the surface to be tilted, and to reflect or scatter incident radiation away from the receiver. Kim (unpublished) observes only a few dB reduction in scattering at normal incidence, while power in later gates is increased due to the occurrence of facets or specular reflectors lying normal to off-nadir incident energy (Dwyer and Godin, 1980). There is, therefore, an inverse relationship between the surface roughness and the gradient of the decay of $\sigma^\circ$ with $\theta$. Information derived from altimeters regarding the variation in $\sigma^\circ$ with incidence angle is a powerful way of characterising the response of the surface to incident altimeter pulses; the only drawback is the restricted range of incidence angles.

5.4 Return pulse characteristics over Arctic sea ice

Several authors (Dwyer and Godin, 1980; Marsh and Martin, 1982; and Rapley et al., 1983) have demonstrated previously that the presence of sea ice within an altimeter footprint causes a number of key signal characteristics. The prominent features of sea ice echoes are the significant increase in signal strength and peakedness of the waveforms, along with the greater variability of both pulse shape and peak power. As shown below, the MIZEX altimetry displays the 'glistening effect' (Robin et al., 1983) or pseudo-specularity associated with sea ice areas. A characteristic ice margin crossing is shown in Figure 5.4 in which altimeter data are compared with statistics on the relative proportions of ice and water.

Altimetric waveforms are presented in a grey-scaled sequence or ZED type display along with 3 descriptive parameters; pulse width, peak amplitude, and waveform integral (ie. pulse energy, or sum of power in range bins). The darker the shade of grey the higher the power value contained in the waveform. Values of ice concentration are derived by digitizing aerial photographs acquired simultaneous to the altimetric data. Before encountering the MIZ margin the altimeter traverses a narrow ice band; then one minute later the aircraft crosses from open water into the MIZ. The response of the altimeter to the presence of ice is identifiable in the increase of peak power, increased variability in pulse width, and variations in pulse energy which appear to have an inverse relationship with ice concentration.
Figure 5.4  Data derived from altimeter waveforms and aerial photography in the MIZ edge region. The upper plot shows the estimated ice concentration and important features. The lowest plot shows the sequence of average waveforms tracked with the interpolation tracker. The ZED-scope type display reads with time delay for each waveform as the y axis and tonal density as increasing power amplitude. The time along track is indicated on the x axis.

5.4.1 Waveform examples from the MIZ

Shortly after the Marginal Ice Zone (MIZ) boundary crossing the aircraft obliquely traverses the dispersed ice margin (see Fig 5.5). An off-ice wind is carrying the outermost smallest floes away from the edge leaving a diffuse ice edge rather than a distinct line. Moving 5 to 10 kilometres into the pack (at the photograph scale of 1:120000) there is 'banding' of smaller floes, and large expanses of open water between the larger floes. The sequence of averaged pulse waveforms in Figure 5.5 shows the evolution of the altimeter returns as it crosses this rapidly varying surface. Apart from the distinct peak, which remains in all this sequence of seven waveforms, the diffuse component becomes more important over the open water areas. The effect is a significant increase of energy in the trailing edge of waveforms, and is caused by waves propagating into the MIZ. Waveform 3 in particular is very distinctive - and similar in many respects to open ocean returns. The only distinguishing feature is the higher peak power. Whereas open sea returns have a typical peak of around 20 bau (Biomation amplitude units) the open water within the MIZ has peaks which are often of the order of 30 bau in magnitude.
Figure 5.5  Mean altimeter waveforms and their corresponding footprints (1-7) crossing a zone of dispersed marginal ice near the ice edge. This nadir photograph covers an area of approximately 13.5 km².
Figure 5.6  Mean altimeter waveforms and their corresponding footprints (1-6) during a traverse of a large ice floe situated 36 km into the MIZ. This nadir photograph covers an area of approximately 13.5 km².
This increase in waveform peak is caused by the sharp contrast in scattering and reflection properties between open sea and water surfaces in the Marginal Ice Zone (MIZ). Recent scatterometry work by Onstott (personal communication; 1984) confirms beliefs that the water between floes has a much higher coefficient of backscatter than does open sea, with increases in $\sigma^0$ of several dB at normal incidence. The suggested cause of this feature is that the sea is calmer in the MIZ. Explanations for this phenomena are given in following sections.

Moving further into the pack the altimeter encounters much larger ice floes. Figure 5.6 shows a series of individual averaged pulse waveforms from footprints which traverse a large floe located at a position approximately 36 km into the pack. Waveforms 2 and 5 are from footprint areas partly filled by ice floes. The proportion of water within them is significant enough to cause a specular return and so the waveforms both have a dominant spike. Peak amplitude is high and close to saturation limit of the receiver. Consequently, there is a small post-cursor to the main body of these waveforms; this being a purely instrumental effect. Normally this is separated from the ‘true’ trailing edge by a dip in power. Waveforms 3 and 4 are from footprint areas completely filled by a large snow covered floe. The scattering properties of snow covered floes result in similar waveform characteristics as those from mixtures of water and sea ice, except that the larger attenuation effects at the snow/air and snow/ice interfaces reduce the returned signal power.

5.4.2 Waveform descriptive parameters

Three parameters are extracted to describe mean pulse waveforms, and are are adapted from Ulander (1985);

1. **RMS Roughness.** This parameter is explained in Chapter 2, and is calculated from the standard deviation of delay times of individual pulses used during construction of mean pulse waveforms.

2. **Leading Edge Slope** ($L_e$). This is calculated by fitting a linear function to five points around the half peak power point in a least squares manner. It is expressed in dB bin$^{-1}$ (where one range bin equals 3.33 ns).

3. **Two Calibrated Coefficients of Backscatter** ($\sigma^0$). Each of the two coefficients expresses the gain (in dB) of an average scatterer on the ice mass surface, assuming that the system point target response is Gaussian as a function of time.

   a) In the first algorithm the backscatter coefficient ($\sigma^0_p$) is calculated using the peak power of the received signal. This gives a measure of backscatter at normal incidence (i.e. from within the first few range bins of the footprint). Since peak value is, in most cases, attained within the first few bins of the waveform it is independent of
the polar scattering diagram of the surface. On the other hand this value is sensitive to saturation effects in high amplitude returns.

(b) An alternative algorithm calculates a weighted value of the coefficient of backscatter ($\sigma_i^0$). This method assumes a non-coherent or 'diffuse' return and utilises the sum or integral of power in the range bins occupied by the returned waveform. It is weighted to account for the antenna beam attenuation pattern. Accurate calculation of the coefficient of backscatter necessitates that the aircraft tilt relative to the surface is small. A pitch, for example, of 1.5° would reduce backscatter by about 0.5 dB, and so only flight sections where pitch and roll and regional slope are small (less than 1.5°) at all times are analysed.

Figure 5.7 and 5.8 show two series of altimeter waveforms from different regions within the MIZ. Figures 5.7(a) and 5.8(a) represent 3-dimensional views of contiguous blocks of mean waveforms stacked in the time domain, and illustrate rapid variations in signal strength and shape due to changes in sea ice characteristics. In each case the corresponding waveform descriptive parameters are displayed beneath in 5.7(b) and 5.8(b). These show how scattering from the water between floes dominates signal amplitude and shape. The parameter traces show that as open water is crossed the value of backscatter at normal incidence ($\sigma_p^0$) exceeds 15 dB in all instances. As the slope distribution and associated wavefield changes the value of $\sigma_i^0$ rises and falls in accordance with increased and decreased diffuse scattering. In most cases of MIZ open water signals typified by the examples at 8.36:6 in Figure 5.6, the value of $\sigma_i^0$ attains values on average 2 or 3 dB less than $\sigma_p^0$. This indicates that the surface is a less than perfect diffuse scatterer, and is not as effective as a fully developed sea outside the MIZ margins which may be regarded as a Lambertian or isotropic scatterer.

In contrast to quasi-specular waveforms, resulting from smooth water surfaces, the parameter traces in both 5.7(b) and 5.8(b) show that as large floes (over 1 km diameter) fill the centre of the footprint, $\sigma_p^0$ decreases from values exceeding 15 dB to less than 12 dB. Although ice surfaces have narrow 3-dB half angles, typically lower than 3°, they have a markedly lower reflexion coefficient, causing the lower signal amplitudes. Another important observation in these cases is that the values of $\sigma_p^0$ and $\sigma_i^0$ fall to two distinct levels over large floes. In a large proportion of these cases they fall to 8 and 3 dB, while in the remainder they fall to 12 and 9 dB, respectively. Over the former of these two it is observed that rms roughness estimates are high, while in the latter rms roughness estimates are low. This would seem to indicate that the altimeter may be able to detect different ice types, due to a combination of material properties and surface roughness: this avenue is investigated further in section 5.6.

Figures 5.7(b) and 5.8(b) illustrate that the RMS roughness estimates range between 1 and 3 metres.
Figure 5.7  Plots of altimeter signal characteristics in the MIZ edge region. (a) shows altimeter waveforms stacked in the time domain along a track section between 8.36:23 and 8.39:3. Variations in waveform shape and changes in surface characteristics are highlighted along the ground swath. (b) shows the corresponding waveform descriptive parameters for the section of data above.
Figure 5.8 Plots of altimeter signal characteristics in the MIZ transition zone. (a) Altimeter signals stacked in the time domain along a flight section between 8.51:5 and 8.53:5. Variations in surface characteristics along the ground track are illustrated and important changes in signal shape are indicated. Note the secondary peak occurring in cases when returns have high amplitudes and narrow peaks. This instrumental ‘ringing’ effect is purely instrumental and occurs when reflected power is high enough to saturate the receiver. (b) Waveform parameters are extracted which correspond with the section of waveforms above, highlighting changes in waveforms due to varying surface conditions.
In cases where large pressure ridges are crossed, for instance, and surface roughness exceeds 2 m, waveform leading edges are stretched. Large peaks in roughness estimates correspond directly with troughs in $L_e$ (i.e. a reduction in leading edge gradient) and decreases in the magnitude of $\sigma_p^2$.

5.4.3 The quasi-specular phenomenon

Peaked pulse waveforms are dominated by the specular component in direct contrast to the 'diffuse' type returns from open sea. Robin et al. (1983) use a simple approach to quantify the relative proportions of both components, which involves considering three types of reflecting surfaces (Robin et al., 1969). This work is applied here in the context of parameters appropriate to the RAL radar altimeter.

5.4.3.1 Plane polished reflector

In this case any departure from a flat surface is small when compared with the radar wavelength. If $P_t$ is the transmitted power, $G_t$ the gain along its axis, $S_r$ the effective absorption cross section of the receiving antenna, $r_0$ the range to the surface; then $P_r$ the received power is given by:

$$\frac{P_r}{P_t} = \frac{R_n G_t S_r}{16\pi r_0^2} \quad \{5.1\}$$

where $R_n$ is the power reflexion coefficient at normal incidence.

5.4.3.2 Perfect diffuse reflector

This type of surface reflects a fraction ($R_d$) of incident radiation according to Lambert's Law. The flux per unit angle reflected in a direction $\Phi$ to the normal to the surface is proportional to $\cos\Phi$, so that the surface has equal brightness when viewed from all directions. For near-normal angles and with certain qualifications regarding polarisation, the power received from the pulse-limited footprint is:

$$\frac{P_r}{P_t} = \frac{R_d G_t S_r c t}{4\pi r_0^3} \quad \{5.2\}$$

where $t$ is the time after the arrival of the leading edge of the pulse. $P_r$ rises linearly with time until time $t_p$, the time after which the reflecting annulus is of approximately constant area, so that $P_r$ is approximately constant in the beam limited footprint. $R_d$ is the diffuse reflexion coefficient.
5.4.3.3 Extended rough surfaces

Beckmann and Spizzichino (1963) consider that backscattering in the direction of normal incidence varies little for surfaces of varying complexity. If a surface is considered upon which slopes are uniformly distributed, up to a maximum value of $\beta_{max}$ (radians), then when the individual facets of the surface are flat to a small fraction of a wavelength over large areas, the problem becomes one of geometrical reflexion, as with the optics of a 'glistening' surface defined by $\Phi_p < \Phi < \beta_{max}$, where $\Phi_p$ is the angle of the pulse limited footprint (Robin et al., 1983). They assume that the power reflected by individual facets is in a random phase relation to its neighbours, and that the average power received is the sum of the individual powers. For a square pulse the waveform rises to an initial maximum given by:

$$\frac{P_r}{P_t} = \frac{R \lambda^2 G_t^2 (ct_p)^{0.5}}{64 \pi^2 r_o^{2.5} \beta_{max}}, \quad \{5.3\}$$

after which the power falls approximately in proportion to $t^{0.5}$ until a time $t_\beta$ when it will fall abruptly to zero; $t_\beta$ is related to $\beta_{max}$ by

$$\beta_{max} = (2ct_\beta/r_o)^{0.5}. \quad \{5.4\}$$

When $\beta_{max} = \Phi_p$, $t_\beta = t_p$, and since $S_r = G_t \lambda^2/4\pi$, we have from equation 5.3 and 5.4

$$\frac{P_r}{P_t} = \frac{1}{\sqrt{2}} \frac{R g G_t S_r}{16 \pi r_o^2}. \quad \{5.5\}$$

This differs from equation 5.1 by only $1/\sqrt{2}$, or about 1 dB. Thus when $\beta_{max} < \Phi_p$ in this case, it appears that it is suitable to treat the maximum returned power as equal to that caused by specular reflexion.

If we consider the relative strengths of diffuse ($P_{rd}$) and glistening or specular returns ($P_{rg}$) from equal proportions areas we can therefore calculate the ratio of energies involved by dividing equation 5.3 by 5.2. Robin et al. derive the following equation;

$$\frac{P_{rg}}{P_{rd}} = \frac{R_g r_o^{0.5}}{R_d 4(ct)^{0.5} \beta_{max}}. \quad \{5.6\}$$

For an aircraft height of 10 km and a value for $t_\beta$ of 100 ns: it is calculated from equation 5.4 that $\beta_{max} = 7.75 \times 10^{-2}$. Using equation 5.6 and the above parameters to find the ratio of energies:

$$\frac{P_{rg}}{P_{rd}} = 58.8 \frac{R_g}{R_d}. \quad \{5.7\}$$

Because the power reflexion coefficient for water is typically 11 dB higher than for most snow/firn covered sea ice surfaces, for equal areas of glistening and diffuse returns:

$$P_{rg}/P_{rd} \approx 6 \times 10^2$$
This means that the peaked component will dominate over the diffuse component if even as little as 1\% of the surface is producing the glistening type of echo. Thus provided that the footprint contains a small area of flat reflective surface exposed to the altimeter pulse, the peaked character will dominate. Since surface slopes due to wave action are reduced as the altimeter moves deeper into the MIZ, intermittent quasi-specularity is expected to occur from areas of smooth sea ice and calm open water. The sequence of averaged pulse waveforms in Figure 5.5 shows the transition from open ocean to sea ice. The response in terms of waveform character is both dramatic and distinctive, and the decay of the trailing edge transforms them very abruptly into the ‘typical’ sea ice pulse waveforms.

5.4.4 On the ‘effective resolution’ over specular surfaces

Typically, the important information contained within the peaked component of RAL waveforms over sea ice regions occurs in the first few range bins after and including peak amplitude. Assuming for the moment a relatively flat smooth surface, and taking account of the ‘system point target response’, the area upon the surface contributing to scattering is considered. Normally over 75 \% of the energy returned from the surface is contained in 8 range bins. In a standard case, the first four of these bins are an integral part of the system point target response, and the surface height distribution, if the surface is undulating. The 4 bins after peak amplitude contain important information on the slope distribution of the surface.

For the RAL instrument the radius \( r_n \) of a range ring on the surface is calculated from the following simple equation (after Rapley et al., 1983):

\[
\tau_n = \sqrt{\frac{n c \tau H}{1 + \frac{H}{A}}} \quad \{5.8\}
\]

where the pulselength (\( \tau \)) is 4 ns, \( H \) is the aircraft height, and \( A \) is the radius of the earth (approximately 6378 km). The nominal radius of the first range ring (ie. \( n=1 \)) is calculated as 100.0 m for an aircraft height of 10 km.

In a similar manner the radii of the Fresnel zones contributing to the coherent returns are calculated. The radius of the first Fresnel zone is 10.5 metres. It is possible therefore, using equation 5.8 to calculate that there are approximately 90 Fresnel zones in each range ring. In section 5.4.1 it is shown that only as little as 1\% of the surface needs to produce phase-coherent echoes for the peaked component in waveforms to dominate. In the case of the range bin containing peak amplitude, where phase-coherent energy dominates over phase non-coherent energy, there are approximately 90 Fresnel zones which may contribute to the signal. Thus, if only 1\% of the surface ‘glistens’, then only one or two Fresnel rings contribute,
to cause quasi-specularity. This result corresponds with findings of Beckmann and Spizzichino (1963) for specular surfaces, and demonstrates that a small circular area of radius 10.5 m (and area 346.4 m$^2$), producing phase-coherent returns, can change the proportions of a RAL altimeter waveform dramatically. It is concluded from this finding that for most flat surfaces encountered in sea ice regions it is unlikely that more than two Fresnel zones contribute to this type of scattering at the nadir point. The corresponding area of two Fresnel zones is 690 m$^2$, with a radius of 14.82 m. This dimension may be regarded as the smallest scale of resolution or the ‘effective resolution’ of the RAL instrument over sea ice areas. Such a resolution enables us to set a lower size limit for features which it is possible to detect within the footprint. Detection of specular reflectors as refrozen leads and polynyas is only possible when they are located directly at nadir.

5.5 Signal responses from floe-filled footprints

The effectiveness of flat water surfaces in dominating the backscatter from the MIZ is clear. But the singular, though less dominant, contribution by scattering from large ice floes should be investigated. MIZ characteristics on the 30th June 1984 were such that the altimeter frequently crossed floes larger than 1 km diameter, providing an opportunity to analyse the response of snowcovered sea ice, without specular contributions from calm water surfaces.

5.5.1 Leading edge distortions over large floes

Walsh et al. (1978) describe the effects of metre-scale surface roughness upon an altimeter pulse incident upon an ocean surface. Under normal circumstances reflexions first occur from the wave crests. The illumination area gradually increases until the trailing edge has reached the bottom of the wave troughs at the nadir point. Provided that the pulse width effective distance (4ns = 1.2m) is small compared with the wave height, the duration of time for the pulse strength to grow to peak power may be many times the duration of time for a smooth planar surface.

In the tracking process for individual RAL pulses a similar stretching of the leading edge over ice floe surfaces is noted. During the construction of mean waveforms the variance of the first return delay times is recorded and the figure translated into an estimate of the rms surface roughness from the standard deviation of the delays (see Chapter 2). This calculation yields a number which is an estimate of the rms height distribution of scatterers, and which corresponds, in most cases, with the degree of stretching of the leading edge attributable to surface roughness.

Pulses which are incident upon ice floe surfaces will be scattered or reflected at normal
of microwave energy at the air/snow interface exceeds 99% (for new snow of less than 250 kg m\(^{-3}\)). Experiments by Onstott (1979) and Parashar et al. (1977), at normal incidence and 13 GHz, conclude that volume scattering is the main mechanism with lower salinity multiyear sea ice. Other work by Ulaby et al. (1984) illustrate that scattering takes place from beneath the surface in this manner, providing that snow covers do not exceed the penetration depth (since this can effectively mask the ice surface). Weak specular reflection may indeed occur at normal incidence, but large proportions of scattering originate from inhomogeneities within the snow volume, such as density variations, from the snow ice interface, and from within the ice volume itself (providing that the ice has rejected its brine). If the extinction within the various layers is minimal then scattering contributions can in theory be made at all levels from zenithal ridge facets at nadir down to the bottom
interface of the ice with sea water. The vertical height distribution of possible scatterers may therefore be dependent not only upon the distribution of surface heights, but also the depth of snow and the penetration depth in the ice (where ice thickness may exceed 5 m).

Leading edges of waveforms may, therefore, be stretched over low salinity multiyear ice to an amount which is the equivalent to the integral of the height distribution of scatterers, spanning the range from the tops of pressure ridges perhaps down as far as the seawater-saturated skeletal layer of ice (associated with the columnar ice crystal growth) at the base of floes. This is a suggested mechanism for the extreme estimates of rms roughness observed for highly ridged snowcovered examples of Arctic multiyear ice. Supporting evidence in the analysis of backscatter records are presented in section 5.5.3.

### 5.5.2 Empirical model for scattering

Various empirical models have been formulated in order to characterise the scattering responses of ground targets. Moore et al. (1980) use a model which is based upon extensive measurements using the Skylab 13.9 GHz scatterometer. They found that the average variation of $\sigma^o$ with incidence angle ($\theta$), was exponential for all terrain types. Furthermore, Ulaby (1980) reports use of a similar model for data collected by another instrument, while Kim (1984) and Onstott (1980) utilise a similar model for scattering from sea ice surfaces. It appears that the following model has a sound experimental basis;

$$\sigma^o = A \exp\left(-\frac{\theta}{\theta_o}\right) \quad \{5.9\}$$

where $A$ determines the absolute level of the curve, $\theta$ is the incidence angle, and $\theta_o$ is the rate of decay of backscatter with angle.

Such an empirical model of exponential form may be transformed into the following straight-line regression formula, which may be fitted to the data in a $\sigma^o$ vs $\theta$ plot;

$$\sigma^o(\theta) = a - \left(\frac{4.34\theta}{\theta_o}\right) \quad \{5.10\}$$

where the constant $a$ is the intercept, or equivalent of backscatter at normal incidence $\sigma^o(0)$, and $-4.34/\theta_o$ is the gradient.

When a measurement is taken over a single target class the dependence of $\sigma^o(0)$ on surface roughness is minimal. Close to the vertical, $\sigma^o$ is observed to vary rapidly with incidence angle (Fung and Eom, 1983). The backscattering coefficient for sea ice between 0 and 15° can generally be approximated by a straight-line and so equation 5.10 fits well to most ice measurements.
Insensitivity of $\sigma^0$ to incidence angle is indicative of increasing surface roughness or volume scatter. Rms roughness estimates of 5 m or more indicate that, in addition to ridging, there is a large degree of penetration. The rate of decay of $\sigma^0$ is thus dependent upon a combination of surface roughness and the proportion of volume scattering.

If snowcover is shallow, surface scatter from beneath dominates. Since the dry snowcover is rarely deep enough to mask the sea ice entirely, the largest proportion of scattering is normally from ice surfaces. Floes with a more reflective surface layer are shown in Chapter 4 to have a higher value of $\sigma^0(0)$, and so a combination of salinity, temperature and surface roughness is likely to control the backscattering coefficient at normal incidence. This may make it possible to distinguish ice types on the basis of scattering intensity $\sigma^0(0)$ and $\theta_o$.

5.5.3 Results

The general principle used in summarising observations made from the plots of backscatter was to find the regression of $\sigma^0$ on $\theta$. This method was used over the range 0 to $10^\circ$ to cover all instances from quasi-specular to diffuse returns for the range of incidence angles subtended within the altimeter's beam limits. Often, it must be noted that half-angles of less than $3^\circ$ are observed, and with the rapid decline in backscatter away from nadir the d.c offset level of 4.8 bau causes an abrupt cutoff in the plots. Despite this drawback, trends may be picked out and the gradient of decay estimated in the regression process.

Data were obtained from large single floes (>1 km diameter). They were derived from reconstructions of the decay in $\sigma^0$ with $\theta$, and achieved using calibrations of the backscattered power values. Where possible, several datasets of $\sigma^0$ and $\theta$ are averaged to construct a mean decay of backscatter with angle for individual floes. Data used in the regressions are expressed in dB. The empirical model was fitted to data from over twenty individual floes of differing properties. In 75% of cases the correlation coefficient was over 0.85, and the standard errors in estimating the gradient or intercept from these fits were more often than not less than 0.2. The exponential fit, therefore, describes the decay of power with $\theta$ adequately.

5.5.3.1 Observed groups

What is apparent from analysis of the regression data is that some similarities appear to exist. Despite the range of surface conditions described in 5.3.1 there appear a number of dominant controls upon observed values. The similarity of their respective backscatter values at normal incidence, and the similarity of gradient in each group suggests that there are four distinct groups of ice types that may be identified from the altimetry. The mean regression lines and their respective equations as plotted in Figure 5.9 are representative of
Figure 5.9  Distinct floe groups identified from least-squares fits to altimeter $\sigma^o$ vs $\theta$ calculations. Parameters derived from fitting equation 5.9 to the data are given for each group, where $A$ is the backscatter coefficient at normal incidence $\sigma^o(0)$ and $\theta_0$ is the e-folding angle (Kim, 1984) or exponential decay parameter.

the characteristic scattering responses of each of the four groups. Notably groups 1 to 3 are very similar, while group 4, in contrast, is distinctly different.

Observations of the supporting aerial photography have been used to a certain extent to validate these results, but it is almost impossible to distinguish ice types on the basis of tonal variations. The main evidence coming from these data is the degree of ridging and rafting, the degree of surface melting and surface roughness, and the structure of the floes. Fortunately, from section 5.3.1, it is likely that there is a strong correlation between the depth of snowcover and the ice type, and perhaps also the degree of surface roughness. Generally speaking, floes with shallow snowcover have suffered the effects of melting to a greater degree, and so floes with patchy snowcover and a visually pitted/rotten surface are more likely to be first year floes. In contrast, multiyear floes are likely to have a uniform snowcover, with a lesser degree of small-scale roughness such as hummocking evident. Ridge heights, on the other hand, are likely to be much higher.

5.5.3.2 Explanations for observed groups

Group 1 is characterised by a high peak value of backscatter of 13 dB, and a typical steep decay, with gradient -1.5 dB/degree. In all group 1 waveforms rms roughness estimates do not exceed 1.4m. Low metre-scale surface roughnesses observed from the photography, in
conjunction with the indication of negligible penetration, supports a hypothesis that these returns originate from first year ice floes. Since, with high salinity ice, signal penetration is normally less than a centimetre, scattering is restricted to the ice surface. Model predictions in Chapter 4 for a bare recrystallised ice surface, simulate these scattering characteristics reasonably well. Floes appear to have undergone extensive surface melting: all display a patchy snowcover. It is suggested, then, that these returns are typical of the first year floes which have visibly undergone surface melt. Flat exposed ice surfaces become recrystallised and desalinised in the upper centimetre layer (Grenfell, 1984). A low degree of surface roughness explains the rapid decay of \( \sigma^0 \), and in all cases little or no pressure ridging is evident from the imagery.

Group 2 has a low variability in peak backscatter values, and display a mean intercept of 11.5 dB. In comparison with group 1 they display a mean gradient which is similar, at -1.6 dB/degree. Lower peak values, 2-3 dB less than group 1, indicate a reduction in reflection coefficient at normal incidence, and thus a change in material properties. This is substantiated by increased variability in rms roughness estimates, and a mean of 2m, as opposed to 1m in the previous group. Model results in Chapter 4 support a hypothesis that such scattering characteristics may be caused by snow covered floes of similar composition to group 1. Backscatter from dry snow covered first year ice floes is typically lower than bare ice, albeit by a few dB only. With a shallow dry snow cover the snow surface roughness contribution is negligible and surface scattering at the ice surface becomes important. Flat areas of the ice surface beneath the snow dominate scattering characteristics, with a peaked response, but power values are lower due to attenuation of power by extinction in the snow layer. Floes are some of the smallest observed, with a mean diameter of 1 km, and have suffered breakup by wave energy more easily than larger surrounding floes, resulting in angular edges. These factors, and tonal variations in the snow cover indicate that these floes are likely to be first year ice, similar in composition to group 1, with a shallow snowcover.

Group 3 is a distinct category of large floes, on the basis of photographic interpretation and scattering characteristics. They have a similar peak value to the previous group (11.4 dB), but a considerably different gradient of -1.0 dB/degree, enough to be regarded as a different group from 1 and 2. Group 3 floes visually fit a description by Maykut (1984) and Ross and Tomchay (1984), of rotted ice. The further south, the higher the frequency of this larger type of floe. In all cases snow cover is sparse and very patchy, and floe structure is revealed to a much greater extent. rms roughness estimates are moderate, ranging between 2 and 4 m, indicating that metre-scale surface roughness or penetration are an important feature in the scattering process. These inferences are supported by the observation of a larger degree of ridging than groups 1 and 2, and a very pitted and hummocked surface indicating melting. Care, however, must be exercised in drawing conclusions regarding ice type from these floe
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observations and scatter regression. There is some evidence that many may be composed of mixtures of first year and multiyear ice types. Burns (personal communication; 1987) has pointed out from SAR imagery that a proportion of large floes have mixed scattering signatures. Surface observations (in section 5.3.2) support SAR interpretations that many of these large floes are a composite of ice types where multiyear pieces of ice are ‘glued’ together by more saline first year ice. It is concluded that this group may include scatter signatures of rotted first year ice, such as those observed by Ross and Tomchay (1984), as well as ice composites. Both have a large degree of metre-scale surface roughness which causes the slow decline in $\sigma^o$ with $\theta$.

Finally, group 4 is also sufficiently distinct to be regarded as a separate category. It forms the largest number of observations, with 7 floes being sampled out of a possible 20. The regression line has a mean intercept of 8.5 dB, markedly lower than in all other cases, and a gradient of -0.37. Such floes scattering characteristics are thus typified by their low peak power and slow decay in backscatter with incidence angle. Most floes have rms roughness estimates exceeding 4 m, and reach a maximum of 7.8 m: these statistics point to a large degree of ridging and penetration. Observations of photography indicate that these floes are generally very large, exceeding 5 km in diameter in most cases. All have a uniform snow cover but surfaces are laced with myriads of pressure ridges. Generally, snow surfaces have an even tone with no indicators of significant melting such as bare ice patches. The surfaces of group 4 floes are not visibly rough between ridges, although snow cover masks ice surface features to a large degree. Floes are also not pitted, hummocked, or rotted as with group 3. Rms roughness estimates certainly indicate penetration, but it would appear difficult to attribute the slow decline in $\sigma^o$ with $\theta$ to surface roughness if little surface scattering is occurring. It is suggested that these floes are composed of much purer ice, allowing deeper penetration and larger volume scatter contributions from both the snow and ice layers. Low gradients observed in their scatter responses matches model results well for volumetric effects from snow covered multiyear ice (with typical snowdepths and ice thicknesses of 20-60 cm and 3-5 m, respectively).
5.6 Performance as a surface roughness sensor

In Chapter 3 and 4 and section 5.5 the influence of surface roughness upon reflection and scattering of electromagnetic radiation clearly demonstrated. The growth and decay of waveforms is modulated by the effects of roughness; and with sea ice/water mixtures to a large extent by the liquid ocean surface roughness. As areas of marginal ice overflown were fairly open, the majority of floes were too small to dominate signals. Except over heavy pack ice, or a continuous sea ice cover, an altimeter senses the water surface roughness. Only floes large enough to fill the footprint cause marked reductions in signal amplitude. The assumption from this point onwards is that marked deviations from typical quasi-specular signals occur due to sea surface conditions alone. In previous sections the response of the instrument in different circumstances has been illustrated. In this brief section physical parameters affecting sea surface roughness are examined, and a number of observations are made and conclusions drawn.

Backscatter cross-section from the surface depends upon reflective properties and the distribution of specular slope facets. The Fresnel reflection coefficient of the air-sea interface at normal incidence has a value between -2.08 and -2.37 dB, depending upon water temperature and salinity. Penetration of a 13.81 GHz pulse is negligible and return signals inherit the signature of the particular wave spectrum. Waveform leading-edge rise time to peak power is affected by the height distribution of the sea surface, whilst the slope distribution of waves influences trailing edge decay times. The ‘direct’ problem of solving the backscatter characteristics given the surface properties is attempted in Chapter 4. Here, the ‘inverse’ problem, to deduce surface properties from observed backscatter characteristics is approached (Boerner et al., 1981).

5.6.1 Ocean waves in the MIZ

Waves, which are considered the major surface roughness elements in this context, possess two scales of roughness. Large, metre-scale ocean waves are predominant but also have smaller ripples or capillaries forming in front of their crests (Crapper, 1957), resulting from a combination of surface tension and wind effects. For this reason, Phillips (1966) describes a sea surface as having a slightly rough surface superimposed upon a long wave structure. Similar ‘composite’ theories in the context of scattering of electromagnetic energy, by Long (1983) include wave facets (i.e. smooth areas on long sea waves) which vary the underlying sea surface slope upon which ripples may or may not be located. A two-scatterer concept may be applied to the altimeter results, where the two principal components are;

a. Gravity Waves
5.6 Performance as a surface roughness sensor

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a. Gravity Waves
b. Wind-dependent fine structure

Bialek's (1966) spectral classification of ocean waves notes 3 categories of wind-generated waves relevant to a near-nadir viewing sensor. Group a includes two of these categories, namely; 'ordinary gravity waves', and 'ultra-gravity waves'. The former is normally generated in the 5-15 s period range, with wavelengths between 0.1 and 100 m and is the usual type experienced on ocean surfaces. The latter typically have a period in the 0.1-1.0 s range but have a combination of surface tension and gravity restoring force. Group b covers the third of these categories, which includes 'ripples' and 'capillary waves'. Surface tension is the main restoring force for such waves with periods less than 0.1 seconds. All three categories identified by Bialek are generated by wind stress. On 30th June 1984 winds of 4 - 5 m s⁻¹ at 10 m (ie. $U_{10}$) are noted by Davidson (1984), Ross and Tomchay (1984), and Guest and Davidson (1987). On the Beaufort Scale this range of windspeed is the equivalent of force 3. On a corresponding sea state scale this gentle breeze may form gravity waves with smooth surfaces, of 0.3-0.6 m amplitude, upon which crests may begin to break. Thus under these circumstances both roughness scales are likely to be present, that is, where the water surface is 'felt' by the breeze. One other component of the long gravity wave portion of the spectrum may be present which is distinct from the 'sea', or local waves generated by wind stress. These are swell waves which have progressed beyond the influence of the generating winds with periods exceeding 15 s and wavelengths between 100 m and 500 m. Sea and swell are generally present in an area at any one time in zones of partial ice cover, but in proportions dependent upon the wave climate and ice conditions.

5.6.1.1 Gravity waves

It is well understood that waves suffer exponential energy decay as they enter the pack, and further, that the short-wave high energy component of the spectrum is preferentially attenuated (Squire, 1984). Studies of wave attenuation in sea ice (Wadhams, 1975; Wadhams et al., 1987) have shown that the primary mechanism for loss of amplitude from the advancing wave is scattering by ice floes, which is a function mainly of the ratio of floe diameter to wavelength. Scattered waves then lose energy by turbulence and wave breaking around floe edges. As a result the largest amplitude losses occur within the outermost section within the MIZ from this and floe-floe collisions. Various theoretical studies, coupled with shipborne wave observations, have investigated exponential decay in the MIZ. Squire and Moore (1980) deployed accelerometers on ice floes in the Bering Sea, and found that for wave periods of 4 s (25 m wavelength) the outer floes of 10 m diameter damp out most wave energy within 5 km. Short gravity waves and ultra-gravity waves are therefore attenuated within a short distance of entering the MIZ outer edge, and the wave spectrum becomes skewed to low frequency, low amplitude long gravity waves. This filtering effect causes a bias in favour of long period swells
which may have been generated hundreds of miles outside the ice edge. Only these waves may propagate more than a few km into the ice. Since the low-frequency end of the wave spectrum makes the largest contribution to the amplitude distribution of the surface (Munk, 1955), it is expected that the surface-height distribution will change accordingly. In addition, because gravity-waves provide smooth tilted facets for specular reflection, it is expected that the angular distribution of reflecting facets will also change with distance into the MIZ.

5.6.1.2 Wind-dependent fine structure

Ripples and Capillaries are important because they form the fine structure of sea surface roughness. They are also the principal mechanism by which the smooth inclined surfaces of long gravity waves are transformed into diffuse scatterers. Short waves are formed by the wind, and according to Kelvin’s definition may exist at wavelengths shorter than 1.7 cm (in clean seawater). The main difference from gravity waves is that they are controlled mainly by the ratio of surface tension to density of water (Cox, 1962). Dissipation effects created by surface tension and wave-floe interaction cause rapid damping, and so these waves are short-lived in the fetch-limited situations found between floes in the MIZ.

Cox (1958) observed short-waves in a laboratory, and using optical methods recorded a bimodal distribution in the wave energy spectrum at low wind speeds. The low frequency peak representing the gravity waves discussed in the previous section has a sharp cutoff at the lowest frequencies caused by fetch-limited conditions. The high frequency peak caused by ripples and capillaries is separated by a trough which represents the frequency of minimum phase velocity of clean seawater (13.5 Hz). As the windspeed increases the gravity peak moves to lower frequencies while the ripple peak moves to higher frequencies and broadens.

Cox and Munk (1954) developed an interesting technique for obtaining slope statistics on the sea surface from aerial photographs of sun glitter on the water. They observed a Gaussian distribution of mean-square surface slope for a fully developed sea, and conclude that it is strongly related to windspeed. Munk (1955) in a subsequent analysis of wind generated ripples clearly showed that it is the high frequency component of the wave spectrum which contributes largely to the mean-squared slope.
5.6.2 Ocean waveform models and altimeter wave measurements

Altimeters have proved successful in measuring waves from space. The ability to extract wind and wave information is based on an interpretation of the slopes of the leading and trailing edges of mean altimeter waveforms. Whilst the significant wave height (SWH) has been determined directly from its effect upon the former, wind speed has been inferred from the trailing edge slope, using a priori assumptions about the nature of wind and wave interaction (Hammond et al., 1977). It has been shown using SEASAT data (Webb, 1981; Brown, 1982) that the accuracy of measuring significant wave heights, for instance, is similar to the accuracy of the buoys used to validate it. Such measurements have been achieved by fitting a theoretical form of the return echo, to the received power at the instrument. The most successful, and widely used model, for generating ocean surface altimeter waveforms originates from work by Moore and Williams (1957), Brown (1977), and Hayne (1980). The theoretical return is derived as a three-fold convolution comprising the probability density function (pdf) of specular points on the sea surface, the flat surface response function (incorporating the effects of the antenna pattern, and angular variation in scattering), and the radar point target response (ie. the transmitted pulse as seen by the receiver). The model, which is reviewed in detail by Ulander (1985) for land applications, has been used widely over the oceans to enable extraction of geophysical parameters such as SWH and windspeed.

![Figure 5.10](image)

Figure 5.10 Predicted pulse shapes using Brown model for variations in (a) rms surface slope, and (b) metre-scale rms surface roughness. All curves have been normalised assuming constant $\sigma^o$ and antenna gain, and a Gaussian variation in $\sigma^o$ with angle.

Examples of waveforms from open water in the MIZ shown in Figure 5.5, may be compared with simulated waveforms calculated using the Brown (1977) model for varying degrees of metre-scale roughness and rms slope in Figure 5.10. It is clear that theoretical fits may be
used to derive values of significant waveheight and surface slope, providing complications such as grease ice are eradicated. In general, sea state within the MIZ is so low that waveforms display a leading edge gradient limited only by the system point target response. Rms roughness values must be of the order of 2 m or less, in Figure 5.5, thereby causing such steep rise times. With low sea states and low windspeeds, the rms surface slope is restricted. Observed waveforms in open water areas within the MIZ display typical delays between peak and half-power point of 33 and 66 ns; corresponding with rms surface slopes between 1 and 3°. In contrast, open ocean waveforms have half power point delays of 150 ns, or more. Estimated rms surface slopes of approximately 7°, correspond with measurements by Gatley and Peckham (1983) of ocean scattering characteristics at low sea states. It is concluded from these observations that open water surfaces within the MIZ on that day had surfaces with rms slope ranging from 0 to 3°, and rms waveheight values less than 2 m. As a direct consequence of the change in wave spectrum and slope distribution during the transition from open ocean into the MIZ power increases by 10 bau (approx 4-5 dB), and delay times between peak and half-power increase by 30 bins or more. It is suggested that statistics of sea surfaces within the MIZ must be studied in greater detail to enable theoretical models to be extended to these situations with a greater degree of confidence.

The main criticism of theoretical models has been that Brown (1977) assumes both the point target response and the pdf of heights of specular points to be Gaussian. Recently, Srokosz (1986) showed that with a modified pdf of specular points, the present model could include the effects of weakly non-linear wave dynamics on the distribution of sea surface elevation (Longuet-Higgins, 1963). This extended theoretical form may be used to estimate other parameters in addition to the significant waveheight, such as the skewness of the sea surface. Lipa and Barrick (1981) and Barrick and Lipa (1985) also suggest that since the transmitted pulse shape is not Gaussian, it is not a simple task to derive information regarding non-linear parameters. However, Challenor et al. (1987) show that providing the transmitted pulse is narrow, the effects induced by assuming a non-Gaussian pulse shape are negligible. So with a degree of modification and tuning, the theoretical model of pulse returns may well be used for returns from within open areas of the MIZ.
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5.6.3 Wind estimates

After initial observations of a relationship between wave slope and windspeed by Cox and Munk (1954) and Munk (1955), various theoretical expressions have been developed to relate $\sigma^o$ to the mean-square surface slope (Barrick, 1974; Hammond et al., 1977), for example:

$$\sigma^o = \frac{|R(0)|^2}{\beta^2} \sec^4 \theta \exp \left[ -\frac{\tan^2 \theta}{\beta^2} \right]$$

{5.11}

where $R(0)$ is the Fresnel reflection coefficient of the air-sea interface at normal incidence, $\beta$ is the rms slope, and $\theta$ is the incidence angle. This relationship is essentially the same as the geometric-optics or Kirchoff scattering model of equation 4.11 by Ulaby et al. (1982).

![Graph showing mean dB vs $\theta$ plot from altimeter waveforms.](image)

**Figure 5.11** Theoretical and observed angular variations in $\sigma^o$ from open water surfaces within the MIZ. The theoretical curve is calculated using equation 4.11, and the mean variation in $\sigma^o$ with $\theta$ is calculated from approximately 400 individual raw waveforms between 8.40:11.79 and 8.40:20.45.

A comparison of theory with a reconstruction of $\sigma^o$ vs $\theta$ in Figure 5.11 illustrates open water angular scattering characteristics, from an average of 400 waveforms, taken from a flight section over a wide lead in diffuse ice of the MIZ eddy. The least squares fit of the theoretical equation in Figure 5.11 gives a value of 0.58 for the reflection coefficient at normal incidence (-2.37 dB), typical of open water surfaces (Robinson, 1985), and an estimated mean square slope ($\beta^2$) of 0.05 rads ($\sim 3.0^\circ$). The plot shows that equations 4.11 and 5.11 may be used with a reasonable degree of accuracy in open pack ice, providing that frazil or grease ice does not interfere with the open water surface.
A variety of empirical relationships have been formulated to derive the windspeed from the calculated mean-squared surface slope (Cox and Munk, 1954). More recently, algorithms have been formulated by Schroeder et al. (1982), Wentz et al. (1982), and Schroeder et al. (1985) specifically for scatterometer measurements. However, unfortunately these are inappropriate in the context of the altimeter because of uncertainties in the measurement of backscatter with viewing angle, and the restriction to angles less than the beam limits. More effective is a calibration of backscatter at normal incidence $\sigma^0(0)$ with windspeed. Recent techniques of measurement have been refined with the use of coincident GEOS-3 and SEASAT altimetry data and buoy measured windspeeds. Brown (1978) suggests that $\sigma^0(0)$ is sensitive to the mean-square slope of a low-pass filtered replica of the true ocean surface. He bases the empirical relationship upon the Phillips (1977) spectral model, in which the mean-squared slope of the filtered surface varies logarithmically with the windspeed. The relationship derived was encouraging, and supported to a certain extent by previous measurements by Cox and Munk (1954).

![Figure 5.12](image)

**Figure 5.12** A plot of the $\sigma^0(0)$ to wind speed conversion algorithm derived by Fedor and Brown (1982) from GEOS-3 comparisons with buoy data. The dotted lines represent the effects of a ±0.25 dB change in $\sigma^0(0)$.

Using this empirical model Brown (1979) was able to correlate $\sigma^0(0)$ and windspeed for GEOS-3 data. More recently Brown et al. (1981) and Fedor and Brown (1982) have improved the model by deriving a three-branch logarithmic relationship, with a small polynomial correction at small windspeeds, between $\sigma^0(0)$ and the windspeed at 10 m height above the surface. A plot of the relationship is displayed in Figure 5.12.

On 30th June, winds were caused by an anticyclone south-west of Svalbard. These weather systems in spring were typically unsustained, and carried cold air masses southwards. Bouts of freezing often resulted from these periods of cyclonic activity in which winds were light ($3 - 5 \text{ m s}^{-1}$), and from $340^\circ$. Calculations of backscatter at nadir, shown in Fig. 5.7b and 5.8b, demonstrate that the relationship shown in Figure 5.12, from the empirical
model of Fedor and Brown (1982), works well, predicting light winds between 0 and 5 m s\(^{-1}\) over the MIZ on that day. The fact that these predictions correspond well with surface observations (Guest and Davidson, 1987) is encouraging, since it would appear that wind estimate algorithms may work equally well on future satellites in open pack areas. A rationale behind observing winds within the MIZ is in the study of patterns of wind stress and the resulting effect upon the ice cover and dynamics. Forthcoming satellite altimeters will be able to make a contribution in this area with suitable calibrations of established algorithms. Possible effects of sea ice in biasing the ocean wave spectrum should be investigated so that assumptions regarding an ocean spectrum in equilibrium are not necessary.

### 5.6.4 Wave penetration into the MIZ

A model is schematized in Figure 5.13 illustrating the typical succession of altimeter waveform shape during the transition from open ocean into the MIZ. Importantly, it shows how pulse shapes and amplitudes change as the wave spectrum is modified by the attenuation effects discussed in section 5.6.1. These waveform changes are attributed by Rapley (1984) to the continuous change in surface height pdf and slope distribution of scattering elements, and the increasing specularity of the surface with distance into the MIZ.

![Figure 5.13](image)

**Figure 5.13** Schematic illustration of the dependency of waveform leading edge slope and trailing edge decay upon the height distribution of scatterers and surface slopes respectively. The diagram shows the altimeter encountering different wave and ice conditions as it flies from open ocean into the MIZ.
Inferences about the ocean surface slope distribution may be made from the reconstructions of $\sigma^o$ with $\theta$ over the MIZ. The main assumption is that returns recorded in successively later bins originate from specular facets upon the surface at successively larger distances away from the nadir point upon the surface. To facilitate specular reflection from points on the surface, at increasing distances, there must be slope facets inclined at an angle equivalent to the incidence angle (measured from the normal). For specular reflection the maximum duration of a pulse is determined by the maximum encountered facet inclination upon waves calculated from a simple geometric relationship, where:

$$\theta \simeq \sqrt{\frac{2c}{\tau H}} \tag{5.12}$$

For an aircraft height ($H$) of 10 km, and a received waveform duration ($\tau$) of 100 ns a maximum wave inclination of 0.08 rads, or 4.4° is calculated.

![Figure 5.14](image)

**Figure 5.14** Frequency distributions of slope facets observed by the altimeter, calculated (a) outside the ice edge, (b) 20 km into the MIZ, (c) 30 km into the MIZ, and (d) 40 km inside the MIZ edge.

Five contiguous sections of a flightline into the MIZ are represented by the statistics plotted in Figure 5.14. The curves represent the frequency distribution of backscattering slopes sampled from groups of over 100 mean waveforms. Although the angular range is limited by the range window delay span, scattering is observed out to angles of 6° or more, indicating that the maximum wave slopes exceed this limit in two wavefields sampled. Curve a represents the ocean wave-spectrum outside the ice edge, with an almost equal probability of observing backscattering facets inclined at angles across the 0 - 6° range. From equation 5.12 inclinations of 6° or more cause durations of 50 bins or 150 ns, and thus the ‘diffuse’ signals typical of open ocean. Curve b illustrates the situation just inside the ice edge, and curves c, and d are from samples at successively larger distances into the MIZ. The statistics go some
way to reconstructing part of the total distribution of wave slopes, and indicate components found in each location. From equation 5.12, a maximum slope of 4°, for instance, corresponds with waveforms of 25 bin, or 80 ns duration. From theoretical predictions in Figure 5.10 this indicates a mean rms slope of approximately 3° in the innermost regions of the MIZ.

A comparison of the observed frequency distributions in a and d (ie. from open ocean and 70 km into the MIZ) reveals that open ocean waves have surface slope components which cover the whole observed range of inclinations. Equal probability of observing different slopes in this range indicates a well-developed sea outside the ice edge. In contrast, the section of flight farthest into the MIZ has a considerably different wave-spectrum. Short gravity waves have by this stage been attenuated, resulting low wave inclinations of only 2° in places, and thus rms slopes between 0.5 and 1.0°. The effect is to increase the frequency of observing near-horizontal wave facets, with a commensurate decrease in observations of higher gradient wave elements.

A continuous reduction in the interrelated slope and height distribution of the ocean surface as a function of distance is consistent with the results of Wadhams (1978), Squire and Moore (1980), Wadhams et al. (1986), and Wadhams et al. (1987). Directional wave spectra and wave attenuation measurements have shown that pack ice attenuates waves differentially, with the high frequency end of the spectrum suffering the largest reduction in energy, even by ice bands outside the ice edge. Although short period wind generated sea may be observed within the MIZ, the likelihood of fetch-limited regrowth of waves becomes less likely with increasing ice concentration.

5.6.5 SWH estimation

Previous observations of the propagation of waves into pack ice have been limited in their extent both in time and space. Interaction of waves and sea ice within the MIZ is of importance since, as explained in section 5.6.1, it controls the size distribution of floes, and thus controls the mechanics, dynamics, and the thermodynamics of ice growth, decay, and breakup. In addition this has consequences where atmospheric-oceanic heat fluxes are concerned, since they influence regional climate. The possibility of making valid wave height measurements within pack ice cover must, therefore, be investigated.

Rapley (1984) describes the use of Seasat onboard (SWH) calculations as a means of investigation of swell penetration in Antarctic pack ice. He suggests that radar altimetry can provide an alternative way of collecting waveheight information in areas of partial ice cover. A similar algorithm has been adapted for use with the RAL altimetry data from the MIZEX campaign to establish the validity of its application in this context and the value of its SWH estimations. It is based upon the Seasat algorithm developed by Miller and Hayne (1972),
and Hayne (1977), and calculates three gates of four varying widths centred upon the leading edge of each respective averaged waveform. The average range gate value within each of the Late and Early gates is calculated and their difference (ie. Late minus Early) is utilised for the SWH estimation. However, firstly the value is scaled and normalized with respect to the integral of waveform power, before being compared with theoretical values stored in a look-up table, to yield an estimate.

Significant Wave Height (SWH) is a statistical property representing the average of the highest 1/3 of waves, sampled over a large (theoretically infinite) area. It is usually defined as $4\sigma_h$, where $\sigma_h$ is the standard deviation of wave heights (Srokosz, 1986). An estimate derived from a limited area will be subject to statistical variability, since it is not fully representative of the total wavefield. The remedy is to average measurements over long portions of the aircraft track. McIntyre et al. (1986) calculate that the minimum value for the integration time, or length of segment of ocean sampled, is a few hundred independent pulses (assuming a constant aircraft speed of 200 m s$^{-1}$), for errors to be reduced to 10% or less. These error estimates, however, rest upon assumptions about the ocean wave spectrum. To assess the importance of such limitations of any given set of data, acquired by the aircraft, there must be simultaneous measurements of the ocean wave spectrum at different places within the MIZ.

Figure 5.15 Significant Wave Height (SWH) estimates using the Seasat algorithm on interpolation tracked waveforms (for a sample size of $N = 100$) shown in the lower fluctuating trace. Signal energy or the sum of power contained in the waveform is displayed as AGC in the upper trace.
The algorithm has been applied to various MIZ transects using mean RAL altimeter waveforms averaged from 200 individual pulse waveforms. Immediate observations show that, in all sections over marginal ice, estimates fluctuate widely; more so than over open ocean. A section of data illustrating the change in SWH estimates and signal strength (AGC) along the flightleg crossing the MIZ edge shown in Figure 5.15. From aerial photography and Plate 5.1 the ice edge is observed to be formed of bands, or 'streamers' of small floes, lying obliquely to the mass of the pack. Before encountering ice the SWH algorithm estimates waves of 3 m amplitude. SWH estimates in Figure 5.15 clearly indicate that the ice has a damping effect upon metre-scale wind-generated gravity waves. Even a 1 km wide band, separated from the main MIZ by some 15 km, registers some effect, with measurements falling during the short crossing period (zeros are returned when $H_{1/3} < 1$ m). Between the ice band and the main MIZ margin at 8.34:29 the algorithm gives estimates of around 3 m. In the immediate period after crossing the MIZ edge SWH values fall rapidly, and the algorithm records values of zero within a few km. Wind generated sea within diffuse ice areas or open water within leads is also picked out by peaks in the SWH trace at intervals along the transect, the largest area of which at 8.36:0.0 returns estimates of 2 and 3 m. Observations of signal energy indicate that with ice/water mixtures in the initial period of diffuse ice, the signal strength rises slightly, before falling dramatically over a section of open water generated by off-ice winds. Thereafter, signal energy becomes more variable, with peaks in energy exceeding 350 units. Powell et al. (1984) also observe increases in signal amplitude and attribute it to attenuation of small wavelength waves, causing highly reflective water surfaces between ice floes. Such surfaces lead to specular reflexion and produce the glistening effect explained in section 5.4.2. High signal strengths from these areas are observed to fall just as rapidly, as the packing density of ice increases and concentration tends to increase. This results in variability in peak amplitude as the aircraft flies over distinct bands of consolidated pack with intercalations of diffuse ice or relatively open water.

In conclusion, the decline in SWH estimates appear to be closely related to the interaction between ice and ice. There is a certain degree of evidence that this in turn is linked to the reduction in height and slope distributions of the surface beneath the altimeter, with distance into the pack (Rapley, 1984). If so, SWH estimates may be used as a crude index of 'surface roughness'. With the supporting experimental observations summized in section 5.6.3 this appears a tenable physical explanation.

The only drawback with the present SEASAT algorithm over the MIZ is that observations of penetration of very low amplitude, and long period swells (wavelengths of the order of one hundred metres or more), require high precision estimates. As well as having to be robust enough to cope with ice and water mixtures within the altimeter footprint, the algorithm estimates must be accurate to a fraction of a metre. Here the Seasat algorithm is inappropriate,
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because low amplitudes, less than about 1 m, result in zero estimates. To achieve any accuracy at low SWH measurements, the errors in tracking and averaging must be reduced to a minimum (less than about 1 ns). Since it has been established by Ulander (1987) that the tracking routine may estimate surface roughness to the nearest 0.1 m, the only limitation is the SWH algorithm itself. Further developments in wave height algorithms are necessary to realise the full potential of altimetry data from marginal ice zones. It must be established how large the integration distance (and thus the number of averaged waveforms) should be within the MIZ, before SWH estimates become accurate. With the prospect of ERS-1 being able to derive wave height information using its onboard SWH algorithm it will then be possible to make quasi-synoptic measurements of swell amplitude, in both the Arctic and Antarctic seas (see Mognard et al., 1979; 1983) and zones of marginal ice within them.

5.6.6 Physical and dynamical causes of roughness variations

Several explanations have been given so far for observed variations in waveform shape in terms of variations in the surface roughness of the liquid ocean in the MIZ. There have been some tentative suggestions for the mechanisms behind them. This brief section gives a brief review of the possible physical and dynamical causes.

Variations in centimetre-scale surface roughness have already been discussed, in the form of wind-generated ripples and capillary waves. Roll (1951) explains the initial formation of such fine structure by the 'bow wave' effect of travelling air pressure disturbances which are convected over the water surfaces by the wind. Phillips (1957) has used a similar theory for ripple generation. Generally, in relatively open water within the pack, wind generated waves cause altimeter waveforms to display the effects of 'diffuse' scattering. However, despite local winds, in a number of areas within the MIZ waveforms become particularly spiked. A number of factors may combine to suppress the wind-dependent 'fine structure';

Meteorological Effects

Variations in sea surface temperature have previously been associated with variations in the stability of air flow above the sea surface. This is known to produce variations in the drag coefficient which determines the surface roughness for a uniform wind field. This mechanism has been proven effective in radar detection of the boundary between water masses of different temperature characteristics (Weissman et al., 1980; Fu and Holt, 1983). It is feasible that mixtures of cold ice-melt water and the warmer North Atlantic waters, being carried north eastwards under the Gulf Stream influence, occur due to instability and vortices at the ice edge. The resulting intermixing of different water masses can have profound effects upon the relative roughnesses of open water areas within the MIZ despite local wind fields, since colder water is known to have a reduced drag coefficient. Davidson (personal communication; 1984)
gives added weight to these suggestions, having observed distinct zones during MIZEX '84, which had what he terms 'stable' and 'unstable' meteorological characteristics. The 'stable' zones are regions where the air is warmer than the water. This leads to reduced transfer of momentum towards the surface within the overlying wind shear. A reduction occurs because the vertical movement of air has to occur against the buoyancy force. Hence, one observes a reduction in capillaries and ripples due to local stress, in stable regions: in unstable zones, wind generated 'fine-structure' is not surpressed. Preliminary results from MIZEX helicopter scatterometry by Onstott support observations that patterns in surface wind stress can be 'seen' at microwave wavelengths as a consequence of the scattering mechanisms explained in Chapter 4. Analogous stable zones are dominated by quasi-specular returns, caused by markedly reduced capillary wave roughness; unstable zones result in lower amplitude returns and more diffuse scattering. This mechanism is thus suspected to be one of the main influences upon patterns of surface roughness and scattering.

**Ice Damping**

In addition to the effects of ice floes covered in section 5.6.1 there is another important effect caused by newly forming ice. The earliest form of sea ice is frazil, and is is defined as a suspension of fine spicules or platelets (discoids) of ice in seawater. Grease ice is the next stage of freezing whereby individual discoids are herded together to form a dense slurry of individual frazil platelets with concentrations by volume in seawater of 20%-40%. Martin and Kauffman (1981) observe plumes of newly forming ice in open water which pile up downwind against the edges of floes due to the effects of wind and wave radiation stresses. Incident wavelengths are observed to be damped out almost completely within 4 to 6 wavelengths. Experiments confirm that grease ice acts as an extremely efficient wave absorber, and Plate A1 and A2 in Appendix A illustrate this point well. The corresponding effect is to make the sea surface extremely 'glassy' and reflected signals are typically saturated, and extremely peaked. Such quasi-specular waveforms are observed in a number of localities, where formation of new ice is confirmed from aerial photography.

**Hydrodynamic Modulation**

Hydrodynamic mechanisms may cause preferential areas of scattering across surfaces of long waves, due to modulation of ripples and capillaries by the phase of the longer period swell or gravity waves (Longuet-Higgins and Stewart, 1964; and Valenzuela, 1979; Alpers, 1983; and Robinson, 1985). Fluid 'orbital' velocity fields associated with the progression of long waves create convergent and divergent velocity patterns at the surface. Such movements result in small scale wave-current interactions and cause the amplitude of short waves and ripples on the crests of long waves to increase, following the passage of a surface convergent zone on the rising edge of each wave. There is a corresponding reduction in ripple energy in wave troughs following passage of surface divergence zones on the backs of waves. Since patterns of increased
and decreased surface roughness are phase-locked to the long wavelength swell, there is more pronounced radar scattering from wave troughs than crests (Fu and Holt, 1983). Dominant scattering from wave troughs may bias statistics extracted from altimetric data regarding wave heights and slopes. These complications must be accounted for in further interpretations of altimeter-derived wave information.

**Internal Waves**

Many observations have been made of the change in surface characteristics of waves generated by wind, when in the presence of internal waves. References range from Ekman (1906), through the papers of Ewing (1950), La Fond (1962), to the work of Apel et al. (1975), in which the observing platform was a satellite. The exact causes of internal waves has not yet been firmly or unambiguously established, although they commonly occur at water mass boundaries or oceanic fronts between low-salinity, low density waters, and higher density saline water (Ekman, 1906; La Fond, 1962). This is expected in the case of MIZ boundaries and associated oceanic fronts, where the ocean has been observed to have a strong two-layer structure. Hughes and Grant (1978) discuss the surface manifestations of internal waves, and indicate that patterns of rough and smooth surfaces are created in patches whose separation is related to the period/wavelength of the internal waves. Phillips (1981) suggests that short waves are positively or negatively strained depending upon whether they propagate in convergent or divergent surface flow sections. In the former case (convergent zones), the spectral energy density of short waves is increased, and in the latter is decreased. This is essentially a hydrodynamic interaction, but is related in this case to the convergence between surface velocities found above the falling limb of the thermocline, behind internal wave crests. Wave-wave interaction is now between internal and surface waves, instead of a two-scale surface wave interaction effect. Apart from straining of wind-generated ripples by the internal wave velocity field, there are other related explanations for a surface signature. A surface-convergence zone can accumulate ice flotsam which might also lead to a smoothed surface rather than a rougher surface, by ice damping (Hühnerfuss et al., 1981; Robinson, 1985). Such surfaces also have the potential to cause glistening.

Internal waves have been observed in radar imagery due to marked changes in ocean surface roughness characteristics (Allan, 1983; Alpers and Salusti, 1983; and Trask and Briscoe, 1983). Importantly, Burns (personal communication; 1987) indicates the presence of internal waves propagating through the MIZ from SAR images obtained by the ERIM CV-580 during the course of the campaign. Caution is necessary in the interpretations of altimetry with respect to these roughness patterns, as there can be no conclusive evidence to suggest that all ‘slicks’, or wave damped surfaces, are related to convergence zones. Although Muench et al. (1983) advance a theory suggesting that regular ice bands may be initiated in flow convergent zones, wind or wave radiation stresses may often prevent these forming. Nonetheless, as SAR
images of the MIZ have indicated the presence of internal waves it appears reasonable to assume that zones of flow convergence and divergence associated with these features may play some part in modulating patterns of surface roughness and thus altimeter response.

5.7 Sea ice algorithm development

Chapters 3 and 4 and previous sections in this Chapter address questions related to the physical basis for many of the observed phenomena in satellite altimetry over sea ice. The work forms the basis for the development of algorithms which will develop sensor output into geophysical units with known accuracy, and which can be used for both applications and scientific research. Other important elements that must be taken into account include sensor calibration, ground data processing operations and user requirements. But, once the understanding of these elements is complete, the task of producing an algorithm to provide geophysical units is straightforward. Unfortunately, a full understanding of these elements is far from perfect, and the task of algorithm development is to make the best job using relationships derived from an imperfect dataset. Nonetheless, algorithms will develop as knowledge of sensor operation improves and as the physics of the remote sensing process are understood.

5.7.1 Ice edge detection and lead and polynya location

Both GEOS-3 and Seasat data have been used to delimit the margin of polar ice packs (Dwyer and Godin, 1980; Rapley, 1984) on the basis of contrasts in backscattering properties of sea ice and open water. On future satellites such as ERS-1, with this capability, it will be possible to delineate the margins of the pack ice, utilising data from successive closely spaced orbits, occurring at, or near, the latitudinal limit of the altimeter.

This ability may be illustrated using RAL altimeter data from the East Greenland Sea. Two simple parameters are extracted. The first is akin to AGC values recorded by Seasat and is calculated from the integration of power values in the waveform (ie. representing the total backscattered energy):

\[ AGC = \sum_{a}^{b} (P - P_{d.c}) \]  \hspace{1cm} \{5.13\}

where \( P \) is the power in a particular range bin, \( P_{d.c} \) is the d.c offset power level (in bau), and \( a \) and \( b \) are range bin positions determined by where a particular threshold power value is exceeded in the signal return. The second is a scaled ratio of the peak power and the AGC. It is an indicator of the distortion of waveforms (INDEX) and is calculated using the peak value of power (PEAK) in the waveform, the AGC value, and a multiplicative factor \( k \) to scale the
index to 1 for standard ocean returns:

\[ INDEX = k \left( \frac{\text{PEAK}}{AGC} \right) \]  \hspace{1cm} (5.14)

Figure 5.16 shows a time series of these two simple parameters characterising the waveform shape during a transect from open ocean into sea ice in the East Greenland Sea. Open water has a constant value of 1, while the ice edge crossing is marked by an abrupt transition from these uniform signals, indicating contamination of the illuminated footprint with sea ice.

\[ \hspace{1cm} \]

**Figure 5.16**  Variation in signal energy and waveform distortion index for a MIZ edge crossing. Ice/water mixtures are picked out by peaks in INDEX. The constant \( k \) is set to a value where \( INDEX = 1 \) over open water areas.

Figure 5.16 also illustrates the considerable variability in signals which is typical of MIZ's. Waveforms constructed from 50 individual pulse returns, transmitted at a PRF of 100 Hz, correspond with an integration period of 1 second. At an aircraft velocity of 200 m s\(^{-1}\), the position of the sea ice boundary can be determined to the nearest 200 m. Such an algorithm may be used in near real-time for rapid dissemination of the sea ice margin location, but may also be used to identify and locate other large regions of open water, such as polynyas and wide leads, to a high degree of accuracy. MIZ margins can be located using this technique with rather more accuracy than NOAA satellite imagery, for example.
Concentration or ice compactness is undoubtedly one of the most significant spatial parameters of MIZ regions. Its value indicates how much ice is present relative to open water. Information on such a property is of paramount importance to scientific research into heat and moisture fluxes between atmosphere and ocean, as well as for the routing of ships and other operational activities taking place in sea ice zones.

One might suspect that concentration would increase with distance into the pack. This simplification is commonly found in most numerical models (Bratchie, 1984). However, the task of modelling becomes one of extreme complexity, since concentration is influenced by a large number of geophysical parameters; for example, wave, current, and wind action causing divergent and convergent stress zones within the pack. Bands of high and low concentration well inside the ice margin are observed on 30th June aerial photography, particularly during the flight section over the large eddy feature. Such variability is caused by the banded structure of the MIZ being disrupted by the action of the vortex. Sections of the ice eddy zone are curled in upon themselves creating a pattern of repeating sets of bands (with respect to concentration and floe size distribution). This being so, the possibility of using the altimeter waveforms, or suitable values derived from them, to yield estimates of concentration is an important area for investigation.

5.7.2.1 Digitizing and image analysis

To complement the altimetry limited sections of cloud-free CV-990 photographic imagery from the 30th June 1984 are available. Suitable subsections of these have been selected using criteria of good variability of ice concentration and floe sizes, along with good altimetric data. From each subsection a systematic sample of images coincident with altimeter footprints are selected and digitized, to yield estimates of floe concentration. Properties of coincident averaged altimeter waveforms are used directly with these figures for correlation purposes.

A technique was developed to obtain accurate statistics on ice floe concentration from the aerial photographic record. Aerial photograph negatives were digitized using a precise automated facility at the Royal Aircraft Establishment, Farnborough. Digital data derived from this process were stored on Computer Compatible Tapes (CCT's). The resolution of the instrument was a fraction of a millimetre, and the accuracy of the process was limited only by the quality of photographic products used.

Digital data stored on CCT's were loaded onto the computer system at the National Remote Sensing Centre, Farnborough. This system enabled the utilisation of a GEMS digital image analyser to process the images, and to manipulate them in such a way that accurate statistics of ice concentration could be derived.
Plates 5.3, 5.4, and 5.5  Plate 5.3 is the original monochromatic image of the altimeter ground swath. It can be displayed on the screen of the GEMS image analyser after being digitized and stored on magnetic tape. Plate 5.4 is an enlarged image subscene magnified and enhanced by contrast stretching techniques. Stretch statistics are overlayed on the image. Plate 5.5 shows the superimposed nominal altimeter footprint. This technique is used to derive image statistics from within the circle only.
Plate 5.3 illustrates a segment of altimeter ground swath, in its original monochromatic form, which is the direct result of the digitizing process. Plate 5.4 demonstrates how individual subsections are enlarged and image contrast is improved by a stretching process. Statistics may be displayed on the screen which indicate the range of brightness levels in both the original image and the ‘contrast stretched’ image. When an optimal result is found, a procedure may be followed to obtain statistics on the relative proportions of ice and water.

Software developed specially by GEMS was implemented to superimpose an altimeter footprint upon the image subsection (see Plate 5.5). Then, to obtain the relative proportions of black and white areas in the image it is necessary to make a classification into groups of ranges of brightness levels or tones. Since the eye can only detect a small number of monochromatic tones it was necessary to classify the image using a ‘density slicing’ technique. The range of 255 brightness levels were sliced into several sections, each of which are assigned an individual colour. The result of an automated classification by this method is displayed in Plate 5.6. In this colour image it is now easier to distinguish between ice and water. Red areas outlined in yellow are ice floes, green areas are small floes and brash, and blue areas are open water.

A number of problems were encountered which made it necessary to use a manual system of density slicing to classify images. In certain negatives the exposure across the frame was uneven, and in others small patches of cloud may have forced erroneous results in the automated slicing process. In addition, where frazil and grease ice, or nilas is present the intermediate grey tones between the dark water and light ice areas resulted in misclassification. The most robust technique, although more time-consuming, was found to be a manual density slice. Most effective schemes of classifying images incorporated only 3 classes, coloured; red (ice), yellow (small floes and brash), and blue (water), in Plate 5.7. This technique enabled boundaries to be established between brightness ranges simply, using visual criteria. Footprints could then be superimposed upon contrast stretched, density sliced images in any position along the ground swath. Plate 5.8 illustrates this final stage, with 3 uncorrelated footprints from which statistics could be obtained. The statistics overlayed upon the screen show the typical bimodal distribution of brightness caused by mixtures of water and ice.
Plates 5.6, 5.7, and 5.8 Plate 5.6 illustrates the result of automated colour classification of an swath subsection using density slicing techniques. Plate 5.7 is a three colour rendition of a manually sliced footprint swath. Red signifies large ice floes, yellow is brash and small floes, and blue areas are open water. Plate 5.8 shows contiguous uncorrelated altimeter footprints superimposed on a manually sliced ground swath enabling derivation of ice concentration statistics.
5.7.2.2 Algorithm investigations

The background to reflexion and scattering in Chapters 3 and 4 indicates that seawater may have a reflexion coefficient some 10-20 times that of first year or multiyear ice. First year ice may have high reflectivities but in general ice and water have scattering coefficients which differ by several dB or more. Preliminary scatterometry results from MIZEX '84 indicate that ice and water may have differences in $\sigma^0(0)$ of 20 dB (Onstott; personal communication, 1985). With these supporting observations it appears reasonable to assume that there is a causal relationship between ice concentration and backscatter or a measure of signal energy, due to the relative proportions of areas with high reflectivities and backscatter coefficient (ie. water) and areas with much lower reflectivities and backscatter coefficients (ie. ice floes).

Figure 5.17 Scattergram of waveform peak power versus ice concentration in the MIZ.

Figure 5.17 is a scatter diagram of estimated concentration versus corresponding peak signal amplitudes. Although there is a wide spread over the range of power and ice concentration the dynamic range of the RAL receiver severely limits high amplitude returns from calm water surfaces. The distinct line formed at 31 bau shows that the instrument operated near the upper limits of its dynamic range during a large proportion of its operation over sea ice transects. In many cases this results in saturation of the receiver (Ulander, 1985). This implies that any investigation of a relationship between signal amplitude and ice characteristics, including saturated returns, is unlikely to represent the true situation. As far as possible, samples have been chosen which minimise waveform distortions or effects caused by instrumental deficiencies, in order to obtain a relationship between concentration and a signal parameter. Signal energy, or the waveform integral, was identified as a suitable candidate since it incorporates the power received in the whole waveform body, and may be calculated simply using equation 5.13 outlined in the previous section. Preliminary results in McIntyre et
al. (1986) have indicated that the waveform integral is indeed related to ice concentration, and Figure 5.4 illustrates this relationship for a short flight section. It is hypothesised, therefore, that there is a direct relationship between the waveform integral and the sea ice concentration.

Before analysis, the dataset had to be reduced in size. A number of data points were extracted where the quality of the negatives was degraded by the problem of uneven exposure across the frame, discussed in the previous subsection. Errors in calculating ice concentration in these excluded sample points are visually estimated at ±20%, upon reanalysis of several badly affected images. The resulting strength of the relationship between the two variables is illustrated in Figure 5.18. Despite the apparent problems encountered in achieving accurate statistics on ice concentrations the datasets have a Pearson product-moment correlation coefficient of -0.74. Standard errors in fitting a regression line are 0.0215 and 3.814, for the gradient and regression constant, respectively. As suspected, therefore, the statistics indicate a moderately strong negative monotonic relationship between the variables.

![Graphs](image1)

**Figure 5.18** The plot illustrates a least squares regression between waveform integral and estimated ice concentration. The resulting regression line equation is given.

**Figure 5.19** An improved correlation and revised regression line equation results with a reduced sample size excluding sample points within 15 km of the ice edge.

A clearer picture emerges when circumstances are reappraised. Effects of waves upon the MIZ waveforms have already been discussed, and it is shown that the wave-spectrum mutually interacts with the type and strength of signals received. Within the first kilometres of entering the ice pack, short gravity waves, likely to interfere with a relationship, are attenuated. In an attempt to clarify the relationship already observed, data points obtained within 15 km of the ice edge are removed from the sample. The new regression in Figure 5.19 is improved, with the
Pearson product-moment correlation coefficient rising to -0.81. The corresponding standard errors in estimating the gradient and regression constant, or intercept, are reduced slightly to 0.020 and 3.51, respectively. In this case the coefficient of determination ($R^2$), expressed as a percentage, gives the proportion of the variance of waveform energy 'explained' by the dependent variable. 65% of the variance in the independent variable is accounted for in the relationship with ice concentration, the remainder being the cause of instrumental deficiencies and other unquantifiable factors.

It is concluded that a relationship, which undoubtedly exists, appears partially dependent upon the prevailing wave climate, and to a certain degree is marred by restrictions imposed by the dynamic range of the instrument. Without the constraints imposed upon this dataset it is suggested, therefore, that with an improved dynamic range the relative difference between the scattering coefficients of water and ice will be more easily distinguishable.

**Figure 5.20** Filtered plot of calibrated backscatter coefficient at normal incidence $\sigma^\circ(0)$ alongside estimated ice concentration for a section of flightline over the MIZ.

The strength of the correlation apparent here is shown in Figure 5.20 in terms of backscatter at normal incidence ($\sigma^\circ(0)$). The calibrated scattering coefficients, derived from waveform peak amplitudes, are plotted alongside ice concentration estimates for a period of several minutes. The rapid fluctuations in power associated with the fading mechanisms discussed in Chapter 2, and a sampling rate 6 times higher than the concentration estimates, are smoothed. When a Butterworth low-pass filter with a cut-off frequency of 0.125 is used to remove the fluctuations, a near perfect one-to-one negative relationship begins to emerge between peaks
and troughs. Although a regression of the two variables only implies a correlation of -0.65, it is clear that the vertical limbs in $\sigma^o$ are restricted, and that a more intensive sample of digitized concentration and improved dynamic range would probably reveal a stronger relationship.

**Algorithm testing**

On the strength of the observed relationships, the hypothesis that there is a direct causal link between sea ice concentration and signal energy is accepted. An algorithm has been defined which may be used in a predictive capacity. Instead of plotting signal energy as the independent variable, a regression line was fitted by a least-squares routine with ice concentration as the independent variable. The regression parameters derived are as follows:

Regression constant (a) = 83.084

Regression coefficient (b) = -0.167

thus;

$$\hat{y} = 83.084 - 0.167x$$  \(\text{(5.15)}\)

where $x$ is the integral AGC, and $\hat{y}$ is the predicted ice concentration. The standard errors in fitting a regression line to the datasets are; 0.021 for the gradient, and 3.70 for the intercept. Analysis of the residual errors indicates a rms residual value of 13.1%.

The algorithm may be used with caution in predictions of ice concentration along altimeter transects into the ice. An rms residual of 13% is reasonable considering the difficulties encountered, and from these studies it may be possible to implement an algorithm to investigate future datasets from marginal ice regions. Present ice concentration algorithms for the Nimbus-7 SMMR sensor, for instance, have equivalent 13% standard errors in retrieval of ice concentration (Comiso and Sullivan, 1986). Comparative studies by Burns et al. (1987), using SAR, dual-frequency passive microwave sensors, and aerial photography from MIZEX '84, show that estimates of ice concentration in the summer MIZ agree to 13% rms error. This suggests that the altimeter may make predictions with equivalent uncertainties, and when used in tandem with existing satellite sensors may make a powerful combination for ice concentration prediction. Additionally, the algorithm suggested for ice margin location may be supported by verification of the proportions of ice and water using the ice concentration algorithm. Simple thresholds may be set to ensure that a concentration of a particular value is met before the coded routine ratifies the ice edge position.

The altimeter is seen to have many advantages over large field-of-view satellite sensors presently used to retrieve ice information. It offers better ground resolution, for instance, than any other instrument used for routine extraction of ice concentration statistics. Though the major drawback of future satellite-borne radar altimeters is the restricted width of the linear transect, the orbit pattern is such that orbital ground swaths are relatively closely spaced at
high altitudes. This feature maximises coverage of the polar regions in a short space of time, and improves temporal resolution.
6.1 Introduction

Ice sheets cover some 10% of the surface land area of the Earth. Their accumulated 36 × 10^6 km^3 of ice comprises 90% of global freshwater reserves (Robin et al., 1983). Such large ice masses are inextricably linked with other aspects of the natural environment through exchanges of mass and energy (Radok, 1978), and changes in volume for example, and are closely related to worldwide sea level and global climate. Because melting of the Greenland and Antarctic ice sheets would together increase sea level by over 65 m, ice volume changes of less than 1% are of great significance (Brooks et al., 1978). Rises in sea level of 5 cm since 1940, documented by Hicks (1978), may be the result in part of changes in the volume of polar ice sheets. Current behaviour of polar land ice masses thus holds important clues to an understanding of contemporary climatic trends.

An increased awareness in the role of the ice in modulating and responding to global climate, and in controlling sea level, has exposed the lack of understanding of ice behaviour (Thomas, 1986). Despite nearly 30 years of intensive fieldwork on the main polar ice sheets it is still not possible to answer the most fundamental glaciological questions. However, detailed observations have revealed two important aspects of ice sheet behaviour which can be studied using microwave remote sensing techniques;

a. Ice sheets are undergoing continual change. Thickening or thinning represents redistributions of mass or mass exchanges with the ocean. Periodic measurements of surface elevation will determine rates of change.

b. Ice sheets have complex feedback relationships with global climate. Cause and effect relationships operate on time-scales which vary from months to millenia. In order to identify and investigate these mechanisms it is necessary to monitor a selection of critical glaciological parameters, including summer melt zones, ice surface elevations, and seaward margins of ice sheets.

Weather over the polar regions can be severe at any time of year; clouds predominate over vast areas, and for several months each winter there is no sunlight. Microwave remote sensing, therefore, with its all-weather, day and night capabilities, plays a critical role in data acquisition and routine surveillance of these areas. Steps may be taken to develop an
understanding through a combination of remote sensing techniques and fieldwork. During the next decade it will be possible to use data collected by a number of forthcoming satellites carrying microwave instruments to assess potential research applications of remotely sensed, and in particular, altimetry data.

Chapter 6 is an investigation of aspects of radar altimetry over land ice surfaces. It is based upon data collected over Nordaustlandet, Svalbard, during a specially organised flight during MIZEX '84 by the CV-990 and the RAL altimeter. This flight took place on 28th June 1984 without remote sensing support and without any organised ground data collection. For this reason supporting data has had to be sought. This chapter briefly describes how radar altimeters can address many of the problems identified. For studies of large ice masses, K_u-band radar altimeters have demonstrated the ability to record precise surface elevations of polar ice sheets (Zwally et al., 1983). Repeated coverage may yield synoptic information on the profiles of major ice sheet and ice cap surfaces, enabling morphological and mass balance studies to be undertaken. Although the main role of an altimeter in providing rapid and accurate surface elevational fixes on ice sheet surfaces is discussed, this technique has already been demonstrated with GEOS-3 and Seasat data (Brooks et al., 1978; Zwally et al., 1983). Simple monitoring of ice sheet elevation by altimeter will not, however, advance knowledge of ice dynamics to the point where it is possible to forecast changes in the ice sheets resulting from climatic changes. Robin et al. (1983) table requirements of several glaciological parameters necessary for a better understanding of ice sheet dynamics. In addition to monitoring dimensional fluctuations they suggest that surface morphology and snowline position are equally important variables. In addition, therefore, to established uses of the altimeter, a number of other potential uses are examined here with RAL airborne altimetry. As well as receiving accurate range data, the RAL altimeter acquires surface information in the record of energy backscattered from the surface. Few investigations have been made of the resulting waveforms, and their relationship to ice surface character. The work shows that prominent surface features, such as coastal boundaries of ice sheets, and seasonal melt and roughness patterns may also be studied with the aid of altimeter data.
6.2 Background

The NASA CV-990 made three flights during MIZEX '84 which traversed Arctic terrestrial ice masses (McIntyre et al., 1986). On 9th June and 1st July the aircraft made two transit flights across Greenland. The flight track on the latter of these two occasions was chosen to coincide as closely as possible with the 'EGIG-line', a profile surveyed by the 1959 Expedition Glaciologique International au Groenland. Data acquired on the 1st July has been analysed in detail by Ulander (1985). Some comparisons are drawn with the results of this study, but otherwise the Greenland data are not discussed. Instead, this chapter is concerned with flight 11, which took place on 28th June 1984 over Nordaustlandet, Svalbard, and with the analysis and interpretation of data records collected from it.

\[\text{Figure 6.1}\] Map of Nordaustlandet ice caps illustrating the full radar altimeter flight track and sections analysed in detail (emboldened). The windows of 4 Landsat image subscenes used as supporting information are outlined as boxes a-d, and the location of Nordaustlandet within Svalbard is inset.
6.2.1 Flightplan: 28th June 1984

The flight schedule on this day was modified to incorporate the flight section over Svalbard, to test the RAL instrument's capabilities over land ice. Figure 6.1 shows the flightpath mosaic over Nordaustlandet, Svalbard. Pattern 'D', as it became known, (Gloersen et al., 1985) consisted of six data legs designed to overlap and retrace previous SPRIL Radio Echo Sounding (RES) flightlines. Several important drainage basins are delineated from which data are analysed in more detail in subsequent sections.

6.2.2 Nordaustlandet: the regional glaciological setting

In terms of terrestrial ice mass areal extent, the Arctic polar region is dominated by the Greenland ice sheet. However, it is accompanied by a number of smaller ice caps, the most important of which are in Svalbard, together covering 11,150 km$^2$ (Drinkwater and Dowdeswell, 1987). These ice caps, located on Nordaustlandet, Svalbard, together represent one of the largest glacierized areas outside the Greenland and Antarctic ice sheets. Viewed as an analogue for large ice masses they have the potential to be used as a laboratory for polar ice sheet research. Their relatively easy access and resulting logistic convenience provide an opportunity to conduct in situ scientific experiments.

Nordaustlandet is the second largest island in the Svalbard archipelago, almost 77% of which is covered by glacier ice. Figure 6.1 indicates ice covered surfaces as white, and all remaining areas as stippled regions. The ice cover consists of two major components: Austfonna, and Vestfonna. Figure 6.2a is a map of Austfonna with contours of ice elevation. Austfonna has an areal extent of 8,105 km$^2$ and comprises two main ice domes, Austdomen and Sørdomen (Dowdeswell and Drewry, 1985; and Dowdeswell et al., 1986). Austdomen has a maximum elevation of 791 m (asl), while the summit of Sørdomen is 683 m in altitude (asl). The two domes are separated by 25 km and a col which is 50 m lower than the Sørdomen summit. Dowdeswell and Drewry propose that since the use of two names is confusing, Austfonna should be applied to the whole ice cap. This name is preferable since it is the more dominant component, and is applied in the following discussion.

Recently there have been a number of studies of the orography of the ice cap from satellite imagery. Dowdeswell (1984), Dowdeswell and Drewry (1985), and Arkhipov et al. (1987) identify several large outlet glaciers which have well defined drainage basins with pronounced margins. The largest, Bråsvellbreen is delineated in Figure 6.1, covering 1100 km$^2$ and 13.7% of Austfonna. Basin 5, in the south-east of Austfonna, is one of several smaller basins also associated with outlet glaciers, and is delineated in Figure 6.1. It comprises 8.3% of the ice cap, with an area of 674 km$^2$. 

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Figure 6.2  Ice elevation maps of Austfonna, Nordaustlandet, after Dowdeswell (1984). (a) Shows a contour map of Austfonna and its surrounding smaller ice caps, Vegafonna and Glitnefonna. The map picks out the two main ice domes Austdomen and Sørdomen, and the regional topography. (b) is an isometric plot of Austfonna spot heights as viewed from the south-west. (c) is an isometric plot of Austfonna viewed from the east.
Figure 6.3  Ice surface elevations of Vestfonna ice cap, after Dowdeswell (1984). The contour interval is 50 m. The CV-990 flightline is traced on the map.

These drainage basins are discussed by Dowdeswell and Drewry (1985) in further detail. The surface morphology of the Austfonna ice cap is represented in Figure 6.2b and 6.2c isometrically; as reproduced from Dowdeswell (1984). These three-dimensional views of the ice cap enable one to visualise the terrain which the altimeter has to track.

Vestfonna, to the west of Nordaustlandet, has an ice cap area of 2,510 km², and is significantly smaller than Austfonna. The ice cap consists of a main divide of inverted 'T' shape, a large proportion of ice draining southwards into Wahlenbergfjorden from an east-west line extended along the bar of the 'T'. 10 km south of the main ice divide, ice streams into well-defined outlet glaciers, namely; Bodleybreen, Aldousbreen, Frazerbreen, and Idunbreen. A map of Vestfonna in Figure 6.3 shows ice surface elevations plotted by Dowdeswell from SPRI airborne radar altimeter data collected during the 1983 RES campaign. An equivalent isometric plot is not available here since flightlines were separated by too large a distance for accurate interpolations of elevations to be made.

The first data on the regime of the ice caps on Nordaustlandet were obtained by Ahlmann in 1931, Glen in 1935-36, and Schytt in 1957-58 (Glen, 1941; Schytt, 1964). The latter two
carried out stationary experiments in the ice divide area of Vestfonna. Schytt (1964) also calculated the total mass balance for all three ice caps for the 1957-58 balance year and found their balance to be positive. Since then various experiments including ice coring (Vaykmyae et al., 1984; and Punning et al., 1985) and airborne radio echo sounding campaigns (Drewry and Liestøl, 1985; and Arkhipov et al., 1987) have led to the conclusion that most of the ice in the two ice caps is close to its pressure melting point, and as such is extremely unstable. By comparing these results with analysis of satellite imagery, Arkhipov et al. (1987) conclude that only in the north-western sector of Austfonna, where there is less likelihood of encountering severely crevassed zones, is ice more stable.

Dowdeswell (1986) uses remote sensing data including aerial photographs and Landsat imagery to measure marginal fluctuations of 22 outlet glaciers for all or parts of the period 1969-1981. All observed outlet glaciers of Austfonna were static or retreating during this period, but a number of Vestfonna outlets advanced for all or part of the time. Notably Bodleybreen exhibited considerable and changing crevassing in association with rapid terminal advance, and is interpreted to be surging. Aldousbreen, Frazerbreen, and Idunbreen may also have undergone periodic surging during their recent trend of advance.

6.2.3 Airborne data sources

Altimetric measurements were made with a nadir-pointing horn antenna on the morning of June 28 between 07.34 hrs and 08.30 hrs. Some 400 track-km of radar altimetric data were acquired over the ice caps of Austfonna and Vestfonna. The data are stored on two CCT’s in digital form, as individual pulse waveforms and precise range delays to the surface beneath the aircraft. Signals were recorded with 196 delay samples, giving a factor of two increase in range window size on sea ice legs discussed in the previous chapter. In the following sections, use of ‘bau’ and ‘bin’ digitizing units considerably simplifies the discussion of data.

Supporting information during the flight was restricted due to a thick cloud cover over the island, which precluded the use of any of the nadir metric camera photography obtained. Only ADDAS parameters recorded for the duration, along with surface radiometric temperatures are available. A selection of the ADDAS parameters concerning the aircraft’s motion, wind direction and windspeed (at 10 km), and infrared surface temperature are plotted as a stripchart in Figure 6.4 for the whole flight section. Passive microwave radiometric data were obtained during the flight by imaging and non-imaging instruments operating at wavelengths from 0.3 to 1.7 cm. Gloersen et al. (1985) conclude from brightness temperatures recorded that the surface layer was close to melting point throughout most of Nordaustlandet. Infrared surface temperature, plotted in Figure 6.4 confirm these observations, remaining fractionally below 0° C for the majority of the flight. The aircraft
maintains an altitude of approximately 10 km, and a speed of around 465 kt ($\approx 250 \text{ m s}^{-1}$).

Figure 6.4 Trace of parameters recorded by the ADDAS during the flight over Nordaustlandet, showing aircraft attitude and motion, windspeed and direction, and remotely sensed infrared surface temperature.

6.2.4 Supporting datasets

Several satellite images have been used as supporting data. Landsat 5 Multispectral Scanner (MSS) imagery from 24 June, and 12 and 17 July 1984 are chosen, since these dates bracket the altimetric dataset and are the nearest days on which partially cloud-free digital scenes are available. Digitally enhanced Landsat subscenes corresponding to each flight section were acquired using digital image analysis facilities at the National Remote Sensing Centre (NRSC), Farnborough. A variety of enhancing techniques were used; contrast stretching to improve image content and filtering to accentuate feature definition. Enlarged image subscenes in Figures 6.5a, 6.5b, 6.5c, 6.5d, provide information on the surface character of the ice caps. Even where slopes are less than 1 or 2°, relatively small changes in ice surface topography are indicated by differences in radiance (Dowdeswell and McIntyre, 1986). Changing ice surface characteristics, for example, snow density, crystal size and water content, will also affect radiance recorded by the MSS.
Figure 6.5  Digitally enhanced Landsat 5 images with flight tracks marked. Each image subscene is located in Figure 6.1 and measures 28 km by 41 km. (a) Bråsvellbreen on 12 July 1984 (Landsat path/row 214/03); (b) Leighbreen on 17 July 1984 (217/02); (c) Basin 5, Austfonna, on 17 July 1984 (217/02); and (d) Vestfonna on 24 June 1984 (216/02).
Digital enhancement of Noroastlandet MSS data recorded on Computer Compatible Tapes (CCTs) reveals considerable detail of the ice cap surface (Drinkwater and Dowdeswell, 1987).

Aerial photographs taken of the ice caps in other years, and field seasons on the ice caps in May 1983 and May 1986, provide further background to the interpretation of altimeter waveforms. Airborne RES data, including both surface and bedrock elevations (Dowdeswell et al., 1986), were important in planning the CV-990 flightlines.

6.3 Noroastlandet: altimetric data summary

Six data legs were completed during the flight, all but one of which contain continuous uninterrupted sections of data. A fixed attenuator setting of 7 dB, combined with the waveform digitizer operating in the ‘video trigger’ mode, resulted in many low power returns not being recorded. Received signals whose amplitude did not surmount a fixed threshold were not detected, and consequently over certain areas with low backscatter intensities triggering is intermittent, making it impossible to use these sections of data to construct representative mean waveforms. The remaining important feature of this flight was that alternate pulses were transmitted by the paraboloid antenna, directed ahead of the aircraft as a look-ahead beam. The effective PRF for nadir transmitted pulses is therefore only 50 Hz, instead of 100 Hz as with sea ice altimetry in the previous section. A preliminary analysis of alternate look-ahead pulses over Noroastlandet is provided by McIntyre et al. (1986).

Data were stored in six individual files, each corresponding with a flight leg, and consisting of between 10,000 and 20,000 individual records. Each record consists of a digitized waveform recorded in histogram form, a delay value representing the time between pulse transmit and Biomation triggering, and a series of flags indicating the quality of the record. The whole data set is summarised in the traces a to f of Figure 6.6. Each section is represented by a profile constructed from the ADDAS radar altimeter delays, upon which the quality of individual RAL altimeter records is superimposed. Sections where triggering of the Biomation digitizer is continuous are indicated by a solid profile. Remaining sections are composed of periods when a proportion of returns are too weak to trigger the digitizer. In these cases, waveforms are recorded intermittently, and sections are identified by dashed lines. Where no signals are recorded at all, or where triggering does not occur for long intervals of time, sections are marked by a dotted line. The ADDAS radar altimeter records are necessary here to show the quality of data along a given profile. In sections where intermittent triggering occurs, the profile traced by the RAL altimeter becomes broken. Despite the fact that the ADDAS altimeter sampling frequency was only 1 Hz, the records are useful in this context to show what terrain the altimeter is crossing, and importantly, in which locations triggering is least frequent.
Figure 6.6  RAL altimetry data quality summary superimposed upon surface profiles constructed from ADDAS radar altimetric data for each of the six flight legs during which data were collected. (a) Bråsvellbreen section in leg 1, as indicated in Figure 6.1 and 6.5a. (b) Gustav Adolf Land flight section (leg 2) in the south-western part of the Austfonna ice cap, over Sørfontna and Vegafontna shown in Figure 6.2a. (c) Long traverse of central Austfontna from Svartknausflya in the south west to Leighbreen in the north west (see Figure 6.1). (d) Leighbreen flight section indicated in Figure 6.1 and illustrated in Figures 6.5b and c. (e) Basin 5 flowline transect on flight leg 5, illustrated in Figure 6.1 and 6.5c. (f) Vestfontna flight section shown in Figure 6.1, and 6.5d.
Flight leg 1

This flight section is illustrated in Figure 6.1; the aircraft flying along a flowline inland up the Bråsvellbreen outlet glacier in the south of Austfonna. A large proportion of recorded data is of good quality, and waveforms are recorded continuously without major breaks for the first 25 km inland. Thereafter, triggering becomes intermittent, with varying proportions of pulses recorded depending upon the scattering properties of the surface. Figure 6.6a illustrates the profile of the surface plotted from the ADDAS records, showing that the regional gradient is only a fraction of a degree.

Flight leg 2

This flight section traverses Etonbreen, the western flank of the Sørfonna component of Austfonna, and Vegafonna in the south-westernmost corner of the island. During periods indicated in Figure 6.6b triggering of the digitizer was particularly infrequent. Large gaps in the records and large amounts of flagged spurious delays render the file unmanageable for analysis of mean waveforms, due to the spacing of the limited sections of good data. Despite infrequent records, good delays may be used to trace the profile of the surface, and in locating where strongest backscatter occurs. The longest section of contiguous records is on the smooth upslope portion of Sørfonna, for a distance of some 4 km.

Flight leg 3

Leg 3 consists of a complete traverse of Austfonna, between Svartknausflya, a non-glacierized region of undulating raised beaches composed of sandy till (Dowdeswell; personal communication, 1985), and Leighbreen in the northernmost part of Austfonna. Figure 6.6c indicates that for the first 20 km over Svartknausflya no pulses are recorded, suggesting that the surface is not a particularly effective scatterer. Immediately the altimeter crosses the ice margin received signals increase markedly in amplitude, indicating that scattering and reflexion from the ice surfaces is far more effective. However, attenuation of pulses due to the extremely steep slopes of Vibeihøgdene, on the southern margin of the ice cap causes intermittent triggering and no recorded pulses for a 3-4 km section over the steepest slopes (> 15°). The most continuous section of good data records occurs after the aircraft crosses the summit of Særdomen and begins to climb to the summit of Austfonna. Good sections on the descending portion of this flightline are interspersed with periods of intermittent triggering, over parts of Leighbreen.

Flight leg 4

Figure 6.6d shows that this data take resulted in a good data set for the first half of the time. Good delays, and large sections of continuous records form the upslope, inbound part of the flight from Kapp Leigh Smith, in the north, obliquely across Leighbreen (see Figure 6.5b). Certain short sections have intermittent delays over the early rougher parts of the ice surface, but continuous triggering reoccurs enabling the remainder to be profiled accurately,
as far as the summit. Thereafter, very few returns trigger the digitizer, despite a very shallow downslope gradient of less than 1° down to Isispynten on the coast, making this remainder of the leg of little use for a waveform analysis or for construction of an accurate profile.

**Flight leg 5**

This was one of the more successful data acquisition legs, and except for the inbound coastal crossing at Hartogbukta (see Figure 6.5c), where scattering of transmitted pulses is affected by the large ice cliff margin. Triggering is almost continuous until the highest point on Austfonna is crossed (see Figure 6.6e). Once the summit is crossed triggering becomes infrequent until the end of the section, where a 4 km long downslope portion has continuous records.

**Flight leg 6**

The final data section was obtained in crossing Vestfonna (see Figure 6.6f). Traverses of Bodleybreen, Aldousbreen, Frazerbreen, and Idunbreen (see Figure 6.5d) cause intermittent triggering, while the smooth dividing ridge sections yield continuous records. The first 50 km of data are useful, until the altimeter encounters Idunbreen. At that point an ADDAS tape change was necessary, resulting in a one-minute downtime. Nonetheless, records become intermittent over Aldousbreen, and so only waveforms recorded between Rjpdalen and Aldousbreen may be used in further investigations.

## 6.4 Terrain effects upon altimetric records

### 6.4.1 Slope-induced effects

Problems caused by loss of lock and faulty tracking are not encountered in the RAL altimetric data set, and so correction and retracking procedures discussed by Martin et al. (1983) are not necessary. Real-time tracking routines are not employed and pulse returns are stored as individual records, something unfeasible for satellite altimeters with limited data storage space between aperiodic telemetry to ground stations. This enables post-processing tracking routines to be optimised for the rapid changes in range delay associated with steep ice sheet margins, and subsequently no records are lost, or waveforms mis-tracked. Nonetheless, despite the added advantages of this technique, certain biases introduced by sloping sections of the surface are still present in the data.

Two major effects caused by surface slopes must be accounted for before analysing altimetry from Nordaustlandet. These are 'range bias' and 'pulse distortion', and both occur over inclined surfaces. The magnitude of these errors with narrow beam satellite altimeters has been demonstrated to be large for seemingly insignificant slopes by Martin et al. (1983) and Brenner et al. (1983), and may seriously affect attempts to reconstruct detailed features of the reflecting surface. These are fundamental limitations inherent in mapping of any irregular
or non-horizontal surface by a radar altimeter. In this short section, both effects are examined in the context of the RAL instrument, before further investigation of altimeter range delays and waveforms.

6.4.1.1 Range bias

Conversion of altimeter range measurements to useful values of ice surface elevation is a difficult task (Martin et al., 1983). The reason is that sloping surfaces introduce biases into the range estimation process. This constant bias, or slope-induced error, occurs because the pulse-limited footprint is upslope and effectively closer to the altimeter than the nadir point directly beneath the instrument. Slope-induced range error is illustrated schematically in Figure 6.7.

![Figure 6.7 Schematic description of slope-induced range error over a planar surface of slope ϕ.](image)

The error ΔH for a single record may be calculated using the following analytical equation;

\[ ΔH = H(1 - cosϕ) \]  

Table 6.1 gives the approximate magnitude of slope induced range bias, and the horizontal displacement of the reflecting surface from nadir. It appears that errors may be serious for slopes in excess of 2°, and become intolerable for slopes over 3°. Horizontal errors are also of interest, since if a height is plotted 348 m out of position it may indicate an error in slope of 1°, and distorts the profile from a straight cross-section.
Table 6.1  Table of the magnitude of range error and horizontal or displacement error for varying surface slopes and a nominal aircraft altitude of 10 km.

<table>
<thead>
<tr>
<th>Slope (degrees)</th>
<th>Range error (m)</th>
<th>Horizontal error (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.0</td>
<td>1.52</td>
<td>174.5</td>
</tr>
<tr>
<td>2.0</td>
<td>6.09</td>
<td>348.8</td>
</tr>
<tr>
<td>3.0</td>
<td>13.7</td>
<td>522.6</td>
</tr>
<tr>
<td>4.0</td>
<td>24.4</td>
<td>695.9</td>
</tr>
<tr>
<td>5.0</td>
<td>38.1</td>
<td>868.2</td>
</tr>
<tr>
<td>6.0</td>
<td>54.8</td>
<td>1039.6</td>
</tr>
</tbody>
</table>

These errors give a guide to the absolute accuracy with which it is necessary to know the aircraft position. If errors of ±350 m were frequent, then it would not be possible to make detailed maps of slopes with greater accuracy, even with an accurately fixed flightpath. With the INS navigation system the approximate limit to the accuracy is ±1 km without corrections for drift in the inertial gyros (Cooper; personal communication, 1986). Profiling accuracy is thus limited to the accuracy of the corrected flightlines and the locational accuracy of the flightline to the nearest kilometre.

Although normal returns can conceivably result from any surface with an inclination which does not exceed the 3 dB half beam-width, most of the surfaces encountered in Nordaustlandet have slopes less than 2° (see Figure 6.2). It may be concluded, therefore, that range errors are unlikely to exceed values of 6 m, and as such are smaller than errors accruing from use of ADDAS pressure altimeter measurements to correct values to sea level. Slope-induced range biases may then effectively be ignored over most of Austfonna, although care is required in constructing accurate profiles in areas where slopes exceed this value. It is clear that for maps of scales smaller than 1:200,000 horizontal errors remain within acceptable limits.

In areas of steep relief, wide-beam altimeters are therefore unable to measure topography with any great accuracy. Nonetheless, range biases due to slope-induced errors are less than figures quoted by Robin (1966) for satellite altimeters, and providing an accurate fix can be made to sea level to a greater accuracy than is presently achieved with the ADDAS pressure altimeter, aircraft microwave altimeters may be used cheaply to provide accurate profiles over areas of ice sheet presently outside the orbital limits of satellite instruments.
6.4.1.2 Pulse distortion

Pulse distortion occurs in connection with range biases because the pulse-limited footprint wanders away from the antenna boresight axis, and region of maximum gain. As a result of antenna beam attenuation, early returns are reduced in amplitude and waveform leading edges become distorted. Figure 6.8a presents a family of idealised diffuse return pulse waveforms for varying inclination of a smooth planar surface. As the surface slope (or equivalent off-nadir pointing angle) increases from 0° the pulse limited footprint wanders further from the centre of the antenna pattern described by Figure B.1 in appendix B, with significant results.

![Figure 6.8](image_url)

**Figure 6.8** Families of simulated waveforms produced using the model of Brown (1977). These idealised waveforms illustrate pulse distortion over inclined surfaces or due to pointing errors, and are normalised assuming constant backscatter at normal incidence. (a) shows the effects of an incline upon pulse returns from a perfect Lambertian surface. (b) shows the same effects for an inclined quasi-specular surface with fixed rms roughness and rms surface slope and a narrow 3 dB backscatter half-angle.

At a surface slope between 3 and 4° the leading edge of the waveform is attenuated by 3 dB, or half its peak power. At steeper angles of attenuation the effect becomes so strong that it is unlikely that backscattered signals are of sufficient amplitude to trigger the Biomation waveform digitizer. Figure 6.8b, illustrates the effect for specular waveforms which Drinkwater and Dowdeswell (1987) identify as being more typical of Nordaustlandet surfaces. In this case, the model waveforms are simulated for a surface with an rms roughness of 2 m, an rms surface slope of 2.5°, and similar variations in surface slope. The corresponding effect is equally important for specularly reflecting surfaces.
6.4.2 Curved or corrugated surfaces

The major limitation upon any altimeter over undulating surfaces is the size of the nominal footprint. The disadvantage of satellite instruments is that the large footprints size and a transmitted pulse-shell, with a radius of curvature of $\sim 800$ km, cannot resolve surface corrugations of greater amplitudes than 20-30 m, and wavelengths shorter than about 16 km. Waveform leading edges are formed from first returns by reflections from undulation summits and the upslope portions of sloping surfaces closest to the aircraft which have surface facets oriented perpendicular to the incident pulse shell. The resulting apparent surface derived from measurements of range represents a smoothed envelope biased slightly above the true surface. Wide-beam airborne altimeters give greater resolution due to the lower operating altitudes, and subsequent smaller radius of curvature of the pulse-shell. Nonetheless, they too suffer effects from undulating surfaces. If the degree of roughness exceeds the pulse-width, waveform distortion discussed in the previous section also occurs. In practice, a realistic surface consists of an assembly of convex and concave components, some of which may be discontinuous, on some spatial scales. Such a distinction may be referred to as local topography, as opposed to surface roughness or regional topography (ie. larger than the BLF). The interaction of incident pulses with local topography causes two effects; smoothing, and focussing.

Figure 6.9  Schematic diagram of the pulse shell incident upon a concave surface of size proportions equivalent to Ericadalen, Gustav Adolf Land (approximately 100 m deep and 2 km wide) depicted in the profile of Figure 6.6b. The radius of curvature of the surface $r_s$ in this case is much less than the aircraft height $H$. As a result energy is returned by backscattering from valley sides oriented perpendicular to incident energy. First returns do not occur from the shaded portion and so the surface profile recorded by the altimeter becomes distorted.
6.4.3 Topographic smoothing and terrain focussing

When traversing narrow troughs, or negative surface features such as the valley at Eri­cadalen, between Austfonna and Vegafonna on Leg 2, the smoothing effect is observed. The situation is depicted in Figure 6.9. First return pulses may never occur from within the shaded portion of the valley, though there is likely to be some scattering from surface facets in the valley bottom at later delays. Indications that the altimeter gives a 'smoothed', rather than the 'true' representation of the surface are borne out by the observation that first return delays specify a surface which is approximately 50 m higher than elevations plotted by Dowdeswell (1984). The dimensions of the feature in Figure 6.9 thus represents the limit to terrain feature resolution for the RAL instrument.

In extreme cases, the surface may consist of a concave dip with radius of curvature equal to the aircraft height. The incident pulse-shell will intercept the surface simultaneously, and the surface impulse response tends to a delta function. This effect is known as terrain focussing (Rapley et al., 1983), and leads to high signal amplitudes and the quasi-specularity more typically associated with highly reflecting surfaces in sea ice areas.

6.4.4 Mis-triggering: 'effective loss-of-lock'

Mis-triggering is caused in circumstances where pulse amplitude is below the threshold required to activate the Biomation waveform digitizer. It is observed in places where backscattering from the surface is not particularly strong, and is linked directly with surfaces over which pulse attenuation occurs. These data drop-outs which the RAL altimeter experiences are synonymous with of loss of lock in Seasat and GEOS-3 altimeter data. The effect is not, however, exclusive to steeply sloping parts of the ice surface. Mis-triggered pulses result in the loss of large sections of data along each data leg. A summary of all sections of data collection legs where mis-triggering is observed to occur is shown in Figure 6.10. Several root causes for intermittent records in the profiles in Figure 6.6a - f are examined here.

6.4.4.1 Receiver dynamic range

The most restrictive feature of the RAL altimeter for the Nordaustlandet flight was the fixed attenuator setting, or fixed receiver gain. For the duration of the experiment it was set to a value of 7 dB (see Table 2.2), limiting the dynamic range of the instrument to a maximum of approximately 12 dB. An upper constraint of 18 dB corresponds with the maximum limit of 31 bau which the digitizing unit could handle, and the lower limit was imposed by the d.c offset, at 4.2 bau, which ranged between about 6 and 8 dB. Both depend upon the aircraft terrain clearance. Thus returns which have backscatter coefficients outside these ranges cannot be recorded using this system. Since snow and ice surfaces scattering at near-normal incidence
have typical values around the lower end of this range, a large number of pulses are not triggered.

6.4.4.2 Slope attenuation

Clearly, the combined effects discussed in section 6.4.1 play a large part in the quality of data recorded, and in the frequency of records. Steep slopes are known to be an important cause of intermittent and blank sections in parts of the data summary profiles indicated in Figure 6.6a to f. Sections of mis-triggering in Figure 6.10 illustrate where slope-induced effects occur.

![Map showing sections of flightline where triggering is discontinuous](image)

**Figure 6.10** Map showing sections of flightline where triggering is discontinuous (marked in bold). The flightline is superimposed upon the surface elevation map to indicate that mis-triggering does not only occur over parts of the surface where contours are closely spaced.

Most notable are sections over Vibehøgdene, the north western margin of Austfonna (see Figure 6.6b), coastal ice cliff areas, the steep slopes of Vegafoøna, and parts of Vestfonna. Cumulative frequency distributions of surface slopes of different angle have been calculated by Dowdeswell (1984) and Dowdeswell and McIntyre (1986) for Nordaustlandet. Four typical frequency distributions are plotted in Figure 6.11 (after Dowdeswell, 1984). Slope angles are generally low with around 70% of the surface of Austfonna less than 1° and 90% less than 1.5°.
Figure 6.11  Frequency distributions describing the occurrence of surface slopes of varying gradient on Nordaustlandet ice caps. The histograms from Dowdeswell (1984) are derived from pressure altimeter and radar altimeter data over these ice caps. A, B, and C are from different parts of Austfonna but display similar characteristics. D in contrast is from Vestfonna and characterises rougher terrain.

Marginal areas of the ice cap have steep slopes in excess of the range depicted in Figures 6.11 a,b, and c, but these amount to less than 1% of the surface of Austfonna. Slopes of Vestfonna are generally steeper than those on Austfonna, as indicated by the profile in 6.6f and the frequency distribution in Figure 6.11d. Nevertheless, over 75% of slopes are accounted for below this limit of 2°.

Figure 6.8b shows that slopes of 3° or more may attenuate pulses by 3 dB, and so even specular surfaces with high coefficients of backscatter may be attenuated by the antenna pattern to levels lower than the trigger threshold. Thus in marginal regions of the ice cap, where slopes reach their highest values, returns with backscatter coefficients at normal incidence of around 10 dB may be attenuated sufficiently to cause mis-triggering.

The main conclusion drawn from this is that mis-triggering is likely to occur on slopes exceeding 3°. This implies that pulse distortion is severe enough over these surfaces for many pulse returns to be attenuated to levels lower than that required by Biomation digitizing unit. However, in cases where steep slopes are not indicated by the profiles in Figure 6.6 and the maps in Figure 6.2 and Figure 6.3, variations in backscatter associated with surface properties
and roughness are thought to be the main cause of lower intensity returns.

6.4.4.3 Reflexion losses

Although data suffer from slope-induced effects and a restricted dynamic range, surface conditions would appear to have the major influence upon the loss of data. The principal discontinuity in the scattering system occurs at the air/snow or air/ice interface. Chapter 3 demonstrates that the maximum and minimum values of the Fresnel reflexion coefficient are -11 dB for solid pure ice, and -18 dB for dry snow of density 300 kgm⁻³; thus the reflexion losses to the forward wave are important, and return waveform amplitudes from backscattered energy are critical in determining triggering.

Chapters 3 and 4 show that the two main factors affecting the intensity of signals reflected from snow and ice surfaces are; the mean density of the medium at the boundary layer with the air, and the amount of free water in the surface layers. Chapter 3 indicates that dry snow is effectively transparent at 13 GHz, and that over 95% of incident energy penetrates. Maximum scattering coefficients which may occur at near-normal incidence are of the order of 6 dB. For surfaces with higher density surface crusts $\sigma^\circ$ may reach 8 dB, and increase above 10 dB for flat wet snow surfaces.

In areas with slopes less than 2° where a significant proportion of data are lost through mis-triggering, it is suggested that surface reflexion loss plays a large part. Dry, low density snow, without ice lenses or marked density interfaces, has a high transmission coefficient, and penetration depths may be as large as several metres. Since volume scattering effects are minimal, in such situations the scattering coefficient comprises a surface scatter component only, and modelled values of the order of 5 dB are too low to be recorded. The only circumstances in which higher backscatter can occur from dry snow surfaces are when marked density variations and seasonal horizons present specular reflectors at shallow depths beneath the surface. Rott et al. (1985) present results from experiments at a similar frequency on dry snow surfaces on Weissfluhjoch glacier in Switzerland. They suggest that typical scattering coefficients at normal incidence are particularly variable, and may range between 1 and 5 dB. Significantly larger signals were observed, but only when snowpacks contained ice lamellae or distinct layers. Thus, dry snow is likely to be the main cause of data loss. One other cause of high reflexion losses is extremely rough areas with a high rms slope, or heavily crevassed zones, for example the surfaces of Etonbreen, Leighbreen, and Bodleybreen.
6.5 Observations of the dynamic topography of Nordaustlandet

Figures 6.5 a-d, the profiles of Figure 6.6a to f, and previous work by Dowdeswell (1984) show that a number of scales of roughness affect incident altimeter pulses over Nordaustlandet. Morphological features which are related to the balance and overall glacier dynamics may be considered as three categories, namely;

(a) large-scale geometry (> 50km)
(b) surface topography (~ 10km)
(c) transient surface features (< 100m)

Category (a) has already been discussed in section 6.2.2, so each remaining category is considered in the context of the RAL airborne altimetry, and discussed in the light of observations made from the data.

6.5.1 Surface topographic observations

Flights were planned to follow previous SPRI Radio Echo Sounding profiles obtained during its 1983 programme as closely as possible. A map of the 1983 RES flightlines is provided in Dowdeswell (1984), showing that flight legs 1 and 3 trace RES flightlines most closely. The ability of the instrument to accurately measure ice surface elevation may be investigated, and the profiles used to investigate topographic features (Brooks and Norcross, 1984).

6.5.1.1 Previous altimeter profiling studies

Glaciologic surveying of terrestrial ice mass topography by radar altimeter was first proposed by Robin (1966), although the intended frequency was in the MHz range. The advent of more precise satellite microwave radar altimeters, despite being optimised for ocean surface measurements, has enabled accurate measurements of ice sheet topography. Zwally (1975) suggested that these more accurate altimeters should be used for synoptic monitoring of ice sheets at 5-10 year intervals, since it is not feasible to go into the field and measure dimensional changes over large areas by normal surveying methods. Such studies will reveal changes attributable to fluctuations in their regime or mass balance.

Brooks et al. (1978) first validated the technique of reconstructing surface topographic profiles from satellite altimeter data using GEOS-3 records from Greenland where rough or steeply inclined ice surfaces were not encountered and problems of tracking were minimal. An accuracy of ~ 2 m was sufficient to establish baseline elevation maps of the southern Greenland ice sheet.
Brooks (1981) discusses the initial performance of Seasat over Greenland and Antarctica, identifying areas of steep slopes where tracking errors and loss of lock occur. Subsequent analyses of these data were undertaken by Brenner and Martin (1982), Martin et al. (1983), Brenner et al. (1983), Zwally et al. (1983), and Zwally et al. (1987). These have shown that over 600,000 useful records were obtained over the Greenland and Antarctic ice sheets, and that biases introduced by the non-standard surfaces can be corrected. As a result, preliminary topographic maps have been constructed for parts of both continental ice masses (Zwally et al., 1983). Final accuracy of absolute elevations is verified by Zwally et al. to be approximately 2 m, or better, over smooth horizontal portions, and approximately 15 m over the steeply sloping and undulating regions.

6.5.1.2 Bråsvellbreen long profile

The Bråsvellbreen flight section, crossing this surging outlet glacier in the southern part of the Nordaustlandet ice cap, is considered here. The surface of this ice stream provides an example with which to examine the agility of the RAL airborne instrument in tracking and profiling ice sheet surfaces. It was chosen since the section has the lowest number of erroneous delays and its surface has a regional slope of a fraction of a degree, thus minimising range bias and pulse attenuation.

![Figure 6.12](image)

**Figure 6.12** Two radar altimeter profiles (corrected to sea level) compared with SPRI radio echo sounding flightline profiles (from Dowdeswell, 1984). The profiles are separated by 100 m to enable comparison. Breaks at regular intervals along the RAL altimeter profiles correspond with interframe gaps in data storage.

Altimetric data from this flight section has been corrected to sea level using the aircraft's
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Altimetric data from this flight section has been corrected to sea level using the aircraft's
digital altimetric data, simultaneously recorded by ADDAS. Data is plotted in Figure 6.12a alongside the RES profile, in which the absolute elevation accuracy is estimated to be 8 m, and the relative error is less than 2 m (Dowdeswell et al., 1986). The profiles agree closely, with the general form of the glacier being picked out quite successfully. Detailed form differs because of some difficulty involved in flying along exactly the same line as that followed during the RES campaign. The relative horizontal displacement of the two flightlines is estimated at an average of a few hundred metres. The precision of the altimeter’s range measurements has been calculated at around 10 cm (McIntyre et al., 1986) while the absolute accuracy of our profile is at best 10 metres, because of corrections to sea-level using the rather more inaccurate ADDAS pressure altitude records.

Errors arise in the case of an airborne platform because of the lack of vertical stability, constant velocity, and a precise flightline. The job of accurate topographic profiling is therefore rather more suited to a satellite. In such a case the ability to track and profile the smooth and gentle slopes of continental ice sheets to accuracies of up to 2m has been amply demonstrated (Zwally et al., 1983). It appears likely that ERS-1’s altimeter, with its so-called ‘ice mode’ (ie. alternative tracking mode with a larger range window or delay span) will provide greater mapping potential. Having an increased size of range window, a factor of 4 larger than Seasat’s, should enable profiling in areas with much rougher and steeper surfaces than those which Seasat succeeded in tracking. This, coupled with a near-polar orbit pattern reaching 82° latitude, promises extended coverage and a valuable synoptic viewing potential of ice sheet topographic detail. Nonetheless, although platform stability is increased, and measurements may be more precise, such a system is dogged by larger range bias errors, problems of tracking, and atmospheric corrections. The advantages of aircraft altimeters are that they may be used without corrections for atmospheric attenuation, and may be used in regions outside the orbital limits of a satellite instrument.

6.5.2 Deviations from theoretical long profiles

Ice flows in the direction of maximum regional slope. Radar altimetry can yield accurate measurements of surface height, and thus provide a key indicator of ice dynamics. Furthermore, since the principal resistance to ice motion is caused by basal shear, the surface slope is a proxy indicator of basal conditions and gross aspects of sub-glacial relief (McIntyre and Drewry, 1984).

6.5.2.1 Bedrock-related surface topography

The profile of Bråsvellbreen plotted in Figure 6.12a, and other flowline profiles from Austfonna and Vestfonna, such as that of Leighbreen (Figure 6.12b), reveal substantial deviations from theoretical equilibrium profiles of quasi-parabolic form. Surface topographic detail highlighted from radar altimetry indicates distortions of uniform ice flow in response to changes
in longitudinal stress and basal shear stress. The generation of surface undulations by flow over bedrock irregularities, for instance, has been discussed by several authors including Robin (1967), Budd (1970), Beitzel (1970), and Budd (1971). Further analyses of measured profiles by Budd and Carter (1971) and Paterson (1981) indicate that the response to bedrock irregularities, in terms of surface undulations, is normally expressed at a dominant wavelength of about 2-6 times the ice thickness. Figure 6.12 shows that the profiles of Bråsvelbreen and Leighbreen have local undulations with wavelengths between 0.5 and 2.0 km. With ice thicknesses which vary between approximately 200 and 500 m, these undulations appear of the correct size magnitude to be related to longitudinal stress gradients in the ice, and thus bedrock irregularities. Such undulations have also been observed by Dowdeswell (1984) and Arkhipov et al. (1987), and appear from radio echo soundings of Nordaustlandet. Several models have been advanced by Paterson (1981) to account for undulations. He suggests that waves generated by viscous-plastic flow over bed irregularities are stationary, whereas kinematic surface waves move downslope, and wind-drift accumulated waves (Black and Budd, 1964) move upslope. Therefore, repeated profiles will enable us to distinguish bedrock waves from other types of surface waves by their stationary character.

Other significant changes from the quasi-parabolic form tend to occur in association with fast flow down outlet glaciers. These exist in Nordaustlandet where ice is funnelled into channels in the main parts of the ice caps, resulting in embayments in the ice cap m-. Spectral analysis of uniform ice flow in response to changes in roughness patterns from altimetry

Dowdeswell (1984) classifies ice surface topography in Nordaustlandet on the basis of formation derived from Landsat images and aircraft altimetry. The spatial extent of rough areas observed in satellite imagery is shown in Figure 6.13, and the implications for the RAL altimetry are clear given the coincidence of these areas and sections of the flight lines. The flightlines marked in Figures 6.5b, c, and d cross visibly rough areas.

Dowdeswell analyses altimeter data from a selection of areas on Austfonna to quantify these local areas of large-scale roughness. Spectral analysis of filtered profiles enables a two-fold classification of surfaces into rough and smooth surfaces. The spectra and surface deviations of the rough areas of the ice cap (depicted in Figure 6.14) indicate that these areas are characterised by undulations of wavelength 3-4.5 km, and 15-25 m amplitude. The smoothest parts of the ice cap surface have amplitudes of less than 10 m and often less than 5 m, with
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Other significant deviations from the quasi-parabolic form tend to occur in association with fast flowing ice streams or outlet glaciers. These exist in Nordaustlandet where ice is funnelled towards the peripheries of the ice caps, resulting in embayments in the ice cap margins, particularly in Vestfonna. Concave profiles along flowlines may also be observed. The example from Bråsvellbreen successfully identifies the zone from which ice was evacuated during the surge activity in 1936-38.

6.5.2.2 Surface roughness patterns from altimetry

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Figure 6.13  Map showing shaded areas of rough ice surface topography on Nordaustlandet as delineated from enhanced Landsat imagery in Figure 6.5 and Dowdeswell (1984).

Figure 6.14  Surface perturbations and spectral power in 4 filtered aircraft altimeter transects over Austfonna (after Dowdeswell, 1984). The upper two traces characterise smooth areas with rms roughness values below 3.2 m. The lower two characterise the shaded regions of rough terrain in Figure 6.13, and both have an rms roughness of 6.2 m.
The energy contained in the surface profiles associated with rough and smooth surfaces varies by two orders of magnitude, and the rms roughness varies from about 1.5-3.5 m in smooth areas to over 6 m for rough ice.

On Austfonna, rough surface topography interfering with altimetry records is confined to an upper section of the Bråsvellbreen flight leg, the surface of Etonbreen for flight leg 2, Leighbreen in the north east, and parts of Basin 5 (see Figure 6.5). Over Vestfonna the altimeter encounters rough topography over the surfaces of ice streams. The surface of Bodleybreen, for instance, is heavily crevassed. Many of these rough areas include parts of glaciers known to surge. Bodleybreen surged during the 1970's, and the surface of Etonbreen and Bråsvellbreen have remained quite rough as a legacy of previous surge activity. The coincidence of shaded sections in Figure 6.10 and rough areas in Figure 6.13 suggests that each of these areas may be used to explain a proportion of mis-triggering of pulses. In certain cases the presence of crevasses may contribute, but low intensity returns are thought to be largely the result of greater scattering of incident energy away from the receiver.

6.5.3 Snow surface microrelief

Metre-scale surface roughness is of critical importance to the scattering process at the surface, and causes stretching of waveform leading edges. Measurements of the amplitude of these features by altimetry are of value in reproducing seasonal patterns of surface windspeeds, and as such are a key element of regional meteorology. Patterns of surface winds and the behaviour of the atmosphere over the ice caps have important controls upon the dynamics, mass balance, and regional and temporal fluctuations in snow accumulation and denudation.

Snow formations discussed in this section have lengths of up to several metres, and amplitudes of up to about 2 m. Although little accurate field-measured information is available on such small-scale features in Svalbard, information provided here is drawn from aerial photographic analysis of the region and published literature.

Two principal processes are responsible for creating these types of roughness. The smallest features are primarily formed by blowing snow. Snowfall is redistributed by this method into irregular waves, ripples, ridges, and hummocks which indicate both the prevailing direction and maximum velocities of boundary-layer winds, as well as snow properties; particularly crystal size, density, and temperature. Such features at the most general level may be considered to be similar to small-scale forms created by blowing sand in arid regions (Cornish, 1914). Dimensions considered typical for these features are from a few centimetres to 1 or 2 m in height (Wright and Priestley, 1922; Tribble, 1964; Kruchinin, 1965; and Ivanov, 1968).

Mather (1962) and Endo and Fujiwara (1973) record a trend for the height of snow microrelief to increase from the centre of the Antarctic ice sheet to a maximum three-quarters
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Mather (1962) and Endo and Fujiwara (1973) record a trend for the height of snow microrelief to increase from the centre of the Antarctic ice sheet to a maximum three-quarters
of the way to the coast. Such a pattern accords with the predominant wind strengths. Winds are weakest on the low gradients in the ice sheet interior where they are insufficient for the development of pronounced snow forms. In strong winds erosional forms are also more likely than depositional ones.

Aerial photographs reveal two types of small-scale feature (Novotny; personal communication, 1986) in Nordaustlandet which have also been extensively mapped and studied in the Antarctic;

(1) Sastrugi
(2) Snow Dunes

Kruchinin (1965) describes the former as an erosional form caused by wind-carving of recently accumulated snow. Amplitudes range from 5-40 cm, lengths vary between 0.2-4 m, and widths are generally 0.2-1.5 m wide. In contrast, snow dunes are a depositional form 18-22 cm high, 3-4 m long, and 6-8 m wide. Both are forms which Kruchinin considers to be typical of early spring, and the former may be expected to occupy between 70 and 80% of the surface, with the latter taking up between 1 and 10%. Observations are made of these features in the interpretation of waveforms in the final section.

6.5.4 Ice margin location by altimetry

Thomas et al. (1983), McIntyre et al. (1986), Drinkwater and Dowdeswell (1987), and Zwally et al. (1987) have all shown that considerable change occurs in altimeter return waveforms as the ice sheet margin is crossed either from fast ice or open water. The sequence of waveforms in Figure 6.15 illustrates the sequence of waveforms as the altimeter crosses the coastal ice margin at Bråsvellbreen.

Seaward ice fronts, floating ice shelves, or glacier margins are generally associated with a step change in surface elevation. Providing the altimeter responds quickly enough to the abrupt changes in surface elevation, location and mapping of the ice front is straightforward. Seasat's altimeter unfortunately reponded too slowly for such changes in range, and failed to retain lock for several seconds or more. Nonetheless, Thomas et al (1983) describe a method for locating the margin of an ice sheet from the records prior to loss of track to within 100-1000 m accuracy. They also suggest that as the altimeter regains lock, after passing over the ice front, distances measured to the ice edge from an oblique crossing may be mapped, giving an indication of the shape of the nearby ice front.

Figure 6.16 shows a sequence of radar altimeter measurements as the instrument passes over a steeply inclined coastal ice margin. Prior to crossing the ice margin, reflexion occurs from the sea ice surface, and waveforms are quasi-specular. The area upon the surface from
Figure 6.15  Series of 9 mean altimeter waveforms during a short section of flight during flight leg 1 over the Bråsvellbreen coastal ice margin.

Figure 6.16  Schematic illustration of the sequence of altimeter reponse to crossing a coastal ice cliff. The diagrams of surface illumination highlight the zones upon the two distinct surfaces, and at different elevations, within which backscattering takes place.

which energy is simultaneously received is indicated by a dark spot, and the area contributing to the trailing edge of the waveform is lightly shaded (Figure 6.16a). As the aircraft passes directly over the ice front (Figure 6.16b) part of the waveform is composed of backscatter from
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Figure 6.16  Schematic illustration of the sequence of altimeter response to crossing a coastal ice cliff. The diagrams of surface illumination highlight the zones upon the two distinct surfaces, and at different elevations, within which backscattering takes place.

which energy is simultaneously received is indicated by a dark spot, and the area contributing to the trailing edge of the waveform is lightly shaded (Figure 6.16a). As the aircraft passes directly over the ice front (Figure 6.16b) part of the waveform is composed of backscatter from
the sea ice surface at the lower level, and the remainder is the result of scattering from the terrestrial ice surface. The area of the pulse-limited footprint (i.e. dark spot) is composed of a semicircular area on the sea ice and a thin semicircular stripe on the ice shelf. Because the beam encounters the surface of the land ice before the lower sea ice level, some backscatter is received before the main peak caused by specular sea ice returns. Because the quasi-specular sea ice returns are of greater intensity, the body of the waveform peak is mainly composed of energy scattered from the lower level on the sea ice surface, despite the fact that the ice front may be nearest. Subsequent returns from both surfaces combine to form the trailing edge of the waveform. When the altimeter is over the ice sheet or ice shelf it measures the new range to the nadir point on the upper surface, and waveforms are shifted to the early part of the range window by the reduction in range clearance (see waveform 3 in Figure 6.15). The backscatter coefficient of the new ice surface may not be as great as the sea ice surface, and so the early peak in the waveform is not as tall as in earlier pulses. A second peak may be distinguished from the main body of the early returns if the range step is large, resulting from the high intensity backscatter from the sea ice surface in the fringe of the beam. However, the geometry of the situation determines the period over which a second peak may exist, and once the sea ice surface becomes obscured by the upper corner of the ice front the late peak disappears. The waveform sequence in Figure 6.15 has already illustrated the typical sequence of waveforms in a sea ice-land ice transition.

The Z-scope type display reproduced in Figure 6.17 corresponds directly with it except that it is constructed from the original untracked, or 'raw' sequence of data. Each individual waveform record is stacked in the time domain, and aligned by a simple threshold. Values of power recorded by the waveform digitizer are represented by pixels of varying intensity; the darkest indicating the highest power values. As the aircraft nears the point of closest approach to the ice front edge, it may be possible with certain geometries to obtain corner reflexion from the foot of the ice front. Quasi-specularity occurs briefly in two places as a strong secondary peak towards the latter part of the range window. The nearer the aircraft becomes to the ice front the closer the spacing of these distinct peaks becomes until the point of closest approach indicated. Thereafter, a second peak re-emerges, due to quasi-specular sea ice returns from the edge of the range window footprint. As the aircraft gets further inland the relative delay of the peaks increases and the intensity of the secondary peak diminishes, due to antenna beam attenuation, until the point where the ice front obscures the sea ice surface in the beam fringe is reached.

The ability of altimeters to accurately locate coastal ice margins is of great importance to mass balance studies. Bråsvellbreen, is of particular importance in this context since it has been known to surge in the recent past. Its formation has been by rapid forward motion and lateral spreading, and so the position of maximum extent is expected to have varied
Figure 6.17  Z-scope display corresponding with the sequence of raw waveforms averaged in Figure 6.15. First returns are aligned, thus eliminating a range delay change when passing over the ice cliff at the point of closest approach. The passage of a second peak into and out of the trailing edge region is clearly demonstrated. The second peak merges with the main peak at the point of closest approach: thereafter it separates again and abruptly disappears.

enormously over the last 50 years. In this way, aircraft altimetry offers a possible solution to synoptic monitoring of glacier and ice sheet coastal margins, and accurate fixing of their locations in time.
6.6 Interpretations of altimeter waveforms from Nordaustlandet

As demonstrated in the previous chapter, waveforms contain information about the geometry of the surface, its scattering and reflexion properties, and its slope distribution. In this section, changes in the waveforms in certain regions are investigated, and a comparison made with corresponding information extracted from selected Landsat 4 imagery and other supporting datasets.

Landsat images provide very useful information on the surface character of the ice sheets in question, in the absence of the aircraft's own photography. Digital enhancement of Landsat MSS computer compatible tapes (CCT's) from several days either side of the flight date reveal considerable detail on the form of the ice mass surface. Observed features were then used to make inferences about the causes of changes in waveform shape, and characteristic patterns/sequences observed in altimetric wavetrains.

6.6.1 Scattering from terrestrial ice at near-normal incidence

The main factors influencing scattering of radar pulses from snow and ice surfaces are outlined before analysis and interpretation of waveforms from the ice caps. Moore and Williams (1957) suggested that mean pulse returns received by an altimeter mainly correspond to incoherent backscattering from a rough surface. If volume scattering is considered negligible, a model mean pulse return may be expressed as a convolution of two parts; the transmitted pulse shape and a function including the antenna beam pattern, ground properties, and range to target. Brown (1977) incorporated and refined these ideas in a model of rough surface scattering (discussed in Chapter 5) by including an explicit expression for the antenna pattern and variation of backscatter with angle. Infrared radiometric surface temperatures plotted in Figure 6.4 indicate that the surface was close to melting point throughout Nordaustlandet. Chapter 3 indicates that under conditions of high free water content in surface snow the resulting dielectric discontinuity between air and snow is large, preventing significant penetration of pulses. The model results of Chapter 4 show that under such circumstances, scattering of the transmitted altimeter pulses occurs almost exclusively at the surface. Even when snow is dry, and reflexion coefficients are as low as 0.05 (c. -13 dB), backscatter at nadir is dominated by surface scattering. Although penetration occurs, interface scattering from ice lenses and marked density variations within the snowpack are thought to be restricted in occurrence. Volumetric effects at near-normal incidence may effectively be disregarded for the majority of normal circumstances encountered on Nordaustlandet. This enables the use of models incorporating rough surface scattering theory to explain variations in waveforms over Nordaustlandet.
Ulander (1985) and Drinkwater and Dowdeswell (1987) discuss the use of Brown's theory to model mean pulse returns over planar ice sheets, with the proviso that its assumptions are only interpreted as first order approximations of surface character. Its main drawback is that ice mass surfaces may be extremely variable across the altimeter footprint. A series of predicted mean RAL altimeter pulse shapes, generated using Brown's model, is shown in Figure 5.10, illustrating the effects of different values of rms surface roughness and rms surface slope. The dependence of scattering upon surface slopes in Figure 5.10a may be quantified from waveforms in terms of the of the 3 dB backscatter half-angle (i.e. the angle from normal incidence to the point at which backscatter coefficient is reduced to half its maximum value, assuming a Gaussian backscatter function). Model waveforms are reproduced in Figure 6.18 for surfaces of varying roughness and 3 dB backscatter half-angle. Increasing rms roughness at the metre-scale reduces peak power and increases leading edge rise time, and increasing backscatter half-angle causes more diffuse returned signals with progressively longer durations. Thus idealised waveforms displayed in Figures 5.10, 6.8, and 6.18 may be used for comparison with data obtained from Nordaustlandet shown in the subsequent analysis.

**Figure 6.18** Predicted pulse shapes for variations in (a) surface roughness, and (b) 3 dB backscatter half-angle using a 7.3 ns pulse duration. All curves are normalised assuming constant backscatter at normal incidence and a Gaussian angular variation in backscatter.

Various approaches to modelling have demonstrated that for comparatively rough glacier ice, the coefficient of backscatter becomes independent of radar frequency and depends on surface rms slope, reflectivity, and incidence angle (Ulaby et al., 1982). For locally smooth surfaces with only gentle undulations, such as snow surfaces, surface scattering may be separated into a coherent (specular) component, and a non-coherent (diffuse) component (Fung,
Properties affecting scattering include:

(i.) Water content
(ii.) Surface roughness
(iii.) Reflectivity
(iv.) Crystal structure
(v.) Density of snowpack

The presence of small amounts of free water in snow is shown in Chapter 3 to affect the loss tangent of the snow significantly. For example, a free water content of 1% by weight will increase the loss tangent from 0.001 to 0.01 in dry snow, thereby substantially reducing signal penetration. The penetration depth through snow is reduced to about one wavelength for free water contents of 3-4% by volume. Temperature variations influence snow crystal metamorphism and alter the texture of the surface, thus affecting scattering. Reflectivity is determined by the permittivity of the medium and controls reflection from and transmission across the air/medium interface. Finally, the crystal structure and density structure, and roughness of dielectric interfaces in the upper layer of the medium are most likely to influence the directionality and strength of scattered energy.

Figure 6.19  Illustration of the major snow and ice facies on an ice sheet surface during the spring-summer transition (after Benson, 1961; and Müller, 1962). Arrows mark places along a surface profile at which major changes in scattering characteristics are likely to be detected.

The surface of an ice cap typically has a great variability, with material properties ranging from low density snow to solid impermeable ice. Figure 6.19 recognises several surface
facies which possess distinctive physical and electromagnetic properties: bare ice, wetted snow, percolation, and dry snow. In the first two, the presence of free water is significant, resulting in surface scattering only. In the latter two, volume scatter is likely, and the effects of ice lenses and percolation features within the snowpack surface layers may influence backscattering. Intuitively, because of differences in material properties between wet snow, dry snow, and bare ice, each is expected to have a different scattering signature. Theoretical scattering signatures of these surfaces presented in Figures 4.10a-f suggest that backscatter signature may be used to discriminate between surface types.

6.6.2 Data analysis

The aims of this section are, first to analyse the nature of radar altimeter waveforms received from Nordaustlandet ice caps and, secondly, to interpret these data with the aid of other information on ice mass elevations and characteristics of the ice cap surface. Three sections of altimetric data emboldened in Figure 6.1 are analysed in detail here to illustrate typical signal variations caused by changes in snow/ice surface characteristics and topography. Information from other sensors is also used to aid interpretation of the altimeter waveforms. Ice surface topographic data from airborne radio echo sounding (RES) of the ice caps are available (Dowdeswell et al., 1986), and the character of the ice surface is also investigated using a combination of aerial photographs and Landsat images acquired close to the date of the flight.

Over the ice sheet signal amplitude varies considerably. Instrument calibration is necessary for the calculation of backscatter coefficients (Ulander, 1985). External calibrations to determine the instrument parameters were undertaken over retro-reflectors located on the Norwegian island of Andøya, and internal calibration was possible during the flight using the instrument’s inbuilt ‘calibration mode’.

Flightline corrections

The aircraft’s flightpath was recorded on magnetic tape by the ADDAS and was derived from the onboard Inertial Navigation System (INS). Firstly, corrections were applied to original records of latitude and longitude to eradicate biases caused by drift of the gyros in the INS mechanism, and the cumulative error was linearly redistributed throughout the flight. Secondly, the flightline was repositioned by fixing its output to the known locations of timed crossings onto the ice caps. Navigational data recorded throughout the flight were used to give an accurate flightline, which was superimposed on satellite images (Figure 6.5a,b,c, and d).
6.6.2.1 Waveform tracking and averaging constraints

Algorithms have been designed by Ulander (1985) which derive mean statistics from RAL altimeter waveforms. Such methods are akin to tracking and averaging routines used to sum and average raw individual pulses on previous altimetric missions (MacArthur, 1978). These act as a data quality check, including sorting routines to guard against spurious data resulting from instrument malfunctions, and geophysically invalid data, becoming incorporated in the analysis.

For an altimeter flying over a variable surface, such as a large ice mass, rapid changes in the form of scattered radar energy may be encountered due to variations in surface geometry and material properties. There is a temptation, therefore, to decrease the number of waveforms averaged in order to minimise the surface area sampled. However, confidence in the mean statistical values is determined by sample size. The control which this has on confidence limits, and the corresponding integration times and along-track integration distances are presented in Table 2.4.

The statistics in Table 2.4 are calculated for a constant aircraft speed of 200 ms\(^{-1}\) and a known pulse repetition frequency of 100 Hz. However, with an unstable moving platform such as an aircraft, these integration distances may only be used as an approximate guide, since the effects of pitch and variable velocity either compress or extend along-track integration distance. Rejecting pulse waveforms during the sorting procedure also extends the sampling interval. For a sample size of 50, the integration distance has been observed to expand, at worst, by 100 m. Normally the use of 100 pulses is statistically adequate to construct a mean waveform. However, for surfaces of high spatial variability it may be necessary to accept increased variance, due to inadequate averaging out of ‘fading’ effects, in order that the terrain characteristics do not change significantly during along-track averaging. Thus, for a mean waveform composed of typically 50 individual pulse waveforms, the 95% confidence limits on the mean power (\(\bar{P}\)) lie between 0.77\(\bar{P}\) and 1.35\(\bar{P}\). This corresponds to a range of approximately 2.5 dB. This range, although large, may still enable us to distinguish between: (a) bare ice and dry snow, which have a possible relative difference in Fresnel reflexion coefficient of up to 7 dB, and (b) wet snow and dry snow, which have a possible relative difference of up to 11 dB in reflexion coefficient (see Chapter 3).

6.6.2.2 Waveform descriptors

Two extra parameters are extracted to describe mean pulse waveforms, in addition to those used in the previous Chapter (after Ulander, 1985):

- **TwoTrailing Edge Attenuation Coefficients** (\(X\)).

Since both coefficients of backscatter are sensitive to pointing errors and surface slope
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Two Trailing Edge Attenuation Coefficients (X).

Since both coefficients of backscatter are sensitive to pointing errors and surface slope
variations, and because they do not take into account variation in backscatter with incidence angle, Ulander (1985) also used two trailing edge attenuation coefficients to measure rates of decay in mean waveforms. They are applicable to waveforms with longer delays where power values are limited by antenna beam attenuation.

(a) The early trailing edge attenuation coefficient \( X_e \) is used to measure attenuation of power in the interval between 6 and 25 range bins after the waveform peak. Signal power falls exponentially, and the following equation is used to calculate the coefficient of exponential decay;

\[ y = a e^{-x} \]  \hspace{1cm} \{6.1\}

where \( y \) is power in units of bau, \( x \) is the coefficient of attenuation \( X_e \), and \( a \) is a constant (300), scaling \( X_e \) to units of \( \mu s^{-1} \).

(b) A late trailing edge coefficient \( X_l \) is calculated in a similar manner for the interval 50 to 89 range bins after the waveform peak.

The early trailing edge attenuation coefficient gives good differentiation between mean returns of short duration, while the late attenuation coefficient gives better separation of longer duration pulse returns. They are chosen to distinguish between surfaces having small 3 dB half-angles (i.e. 1-3°), and surfaces having large 3 dB half-angles (i.e. > 3°). The latter coefficient is also independent of surface height distribution effects on the received signals.

The above parameters are calculated for three sections of the altimeter flight over Austfonna and Vestfonna.

6.6.3 Variations in waveform parameters: Bråsvellbreen

6.6.3.1 Results

Mean waveforms from the long profile of Bråsvellbreen (Fig. 6.1) are illustrated in Figure 6.20, with corresponding waveform parameters in Figure 6.21. Rms roughness is calculated from the standard deviation of pulse delays (Fig. 6.21f). Figures 6.21a to 6.21g reveal three zones, after initial waveform disruption caused by the inbound coastal crossing. Note that the boundaries between ice surface zones discussed here are in reality transitional.

The first identifiable region stretches from approximately 3 to 17 km, and is characterised by the mean returns shown in Figure 6.20a and 6.20b. It is a period of relative stability and \( \sigma_p^2 \) and \( \sigma_r^2 \) show only slight steady increases, from 12 to 14 dB and 6 to 8 dB respectively. \( X_e \) stays fairly constant at around 25 \( \mu s^{-1} \), while an increase of \( X_l \) from 2 to 10 \( \mu s^{-1} \) picks out the extended tails of waveform trailing edges. Leading edge gradient \( L_e \) (Fig. 6.21c) fluctuates
Figure 6.20 Four contiguous mean pulse waveforms taken at five intervals along the Bråsvellbreen flight section (see Figure 6.1) illustrating varying characteristics at a, b, c, d, and e in Figure 6.21.

between 1.5 and 2.0 dB bin$^{-1}$, after having been stretched to 0.5 dB bin$^{-1}$ at the 3 km stage. rms surface roughness varies between 0.5 and 2.0 m (Fig 6.21f).

The second region stretches from 17 to 25 km, and is characterised by the mean waveforms in Figure 6.20c. Peak signal amplitudes reach their maximum, with $\sigma_0^0$ attaining a plateau at around 15 dB. Waveform duration increases to 300 ns resulting in a fall of $X_e$ from 30 to 20 $\mu$s$^{-1}$.

The final region continues from 25 km until the end of the profile, and mean signals received at 31 km inland are displayed in Figure 6.20d. A marked reduction in signal amplitude occurs over the space of 5 km, and the coefficients of backscatter $\sigma_0^0$ and $\sigma_1^0$ fall by 5 dB to 9 dB, and by 6 dB to 4 dB respectively. Additionally, $L_e$ (Fig. 6.21c) falls to minima of below 1.0 dB bin$^{-1}$. More important, however, is the change occurring at around 29 km inland after which both trailing edge attenuation coefficients become markedly more variable. Beyond 35 km inland signal amplitudes fall to their lowest values, and mean returns observed at 38 km inland are displayed in Figure 6.20e. At this point the proportion of spurious delays increases, as many signals are too weak to be detected by the receiver. This situation is characteristic of the central parts of the ice cap.
Figure 6.21 Traces of varying waveform parameters and ice surface profile along Bråsvellbreen flight section. The arrows labelled a to d in part g of the figure represent the locations of the sets of waveforms in Figure 6.20.

6.6.3.2 Interpretation

Bråsvellbreen surged between 1936 and 1938, and is now in the quiescent period between surge activity (Schytt, 1969). It has a regional gradient of less than a degree, thus minimising inaccuracies in backscatter calculations.

Comparison of the mean returns in Figure 6.20 with the model waveforms in Figures 5.10a and 5.10b indicates that over the sections identified progressive changes in waveform geometry occur. Waveforms in Figure 6.20a, 6.20b, and 6.20c illustrate the increasing mean signal amplitudes and durations between 7 and 19 km. Throughout, signals have a specular component which dominates over a diffuse scattered component. As the bulk of each waveform is composed of reflected energy they are termed 'quasi-specular'. Since signals do not result principally from diffuse scattering mechanisms, the two values of $\sigma^o$ (Fig. 6.21a and 6.21b) show a consistent discrepancy, with $\sigma^o_p$ continually greater by 5 dB. Only in places where diffuse scattering becomes dominant will the two values coalesce and follow similar trends. Otherwise, although incorrect as an absolute value of $\sigma^o$, changes in $\sigma^o_p$ indicate variations in scatter contributions from angles off-nadir (Fig. 6.21b). At this point the surface has a narrow 3 dB backscatter half-angle, attaining a value of 3° at 19 km inland. Tonal variations on Landsat imagery indicate that most of the ice surface is snow covered, apart from small
coastal areas where bare ice is visible (Fig. 6.5a). At this time of year the snow depth is likely to be significantly less than 1 m in depth within 25 km of the coast, and surface melting results in high free water content in the upper layers. Snow depths are based on surveys by Scott Polar Research Institute field parties in spring 1983 and 1986. It is suggested that in relatively wet snow areas the main mechanism causing near-specular signals is probably that of reflection from flat surfaces orientated perpendicular to incident pulses. Additional important effects are thought to be caused by reflection from supraglacial streams which cross the lower parts of the wet snow zone. These streams are visible beneath thin cloud on the Landsat scene from 12 July (Fig 6.5a).

Maximum observed signal amplitudes occur between 17 and 25 km, and large 3 dB scattering half-angles (over 3° in places) cause longer duration waveforms (Fig. 6.20c). More dense refrozen or glazed surface crusts, or shallow ice lamellae or lenses in this region (Schytt, 1964; and Arkhipov et al., 1987), explain the continuing dominance of the specular component of waveforms over the diffuse component.

From 25 km onwards signal amplitude falls considerably. This may be a response to moving to progressively higher elevations, where the effects of melting are reduced (Fig. 6.20d). Surface melting and the formation of ice lenses has been observed even on the high crest of Austfonna (Schytt, 1964; Dowdeswell; personal communication, 1986). However, whether or not surface melting had taken place high on Austfonna by late June 1984 is not certain. Trailing edge attenuation remains fairly stable until 30 km, whereupon $X_e$ and $X_l$ begin to fluctuate, in response to 3 dB backscatter half-angles varying between 3 and 7°. Variability in polar scattering characteristics of dry snow surfaces has been observed by Ulander (1985). He derived a similar range of values from altimetry over central Greenland on 1 July. The cause of this variability is not understood, but is most likely to be a combination of small-scale roughness, the reduced effects of snow-grain metamorphism, and increasing volume scattering effects.

In relatively dry snow some penetration may occur, and volume scattering is possible. However, at near-normal incidence surface scattering continues to dominate. If a Gaussian height distribution is assumed then the value of 1.5 dB bin$^{-1}$ for $L_e$ (Fig. 6.21c) at 33 km corresponds with a value of 1 m surface roughness predicted by Brown’s model for a similar leading edge gradient. This matches calculations of rms surface roughness at 33 km (Fig. 6.21f), and is similar to the height of wind-generated snow features such as sastrugi, dunes, and ridges. Snow feature dimensions observed by Ivanov (1968) range from a few centimetres to 2 m amplitude, with long axes from metres to several tens of metres. Beyond 30 km, peaks in rms roughness are observed (Fig. 6.21f) indicating larger scale roughness than aeolian features, (2 and 3 m). These account for some reduction in backscatter at nadir and a reduction in
leading edge gradient. Signals in Figure 6.20e have more rounded peaks and considerable variability in late bins.

6.6.4 Variations in waveform parameters: Basin 5, Austfonna

6.6.4.1 Results

Three zones are identifiable in this section of data; 0-7, 7-25, and 25-33 km (Figure 6.22). Between 0 and 7 km inland $\sigma_p^o$ is consistently the highest value of backscatter by 10 dB (Figure 6.22a and 6.22b). $L_e$ remains steady at 1.3 dB bin$^{-1}$ (Figure 6.22c), but the attenuation coefficients fluctuate; $X_e$ between 0 and 35 $\mu$s$^{-1}$, and $X_I$ between 0 and 15 $\mu$s$^{-1}$. The surface profile in Figure 6.22f has discontinuities indicating periods where the altimeter records spurious delays and mis-triggering occurs.

![Figure 6.22](image)

Figure 6.22  Traces of varying waveform parameters and ice surface profile along Basin 5 flight section, Austfonna.

From 7 to 25 km $L_e$ becomes greater, attaining values comparable to those observed over Bråsvellbreen. There are increases in signal amplitude indicated by peaks in $\sigma_p^o$, but $\sigma_i^o$ rises dramatically by over 10 dB in places following similar trends to $\sigma_p^o$. In contrast, $X_e$ falls to low values between 15 and 20 km, periodically reaching zero. $X_I$ attains values similar to those
given over the section of several kilometres before the inbound coastal crossing indicated on the surface profile (Figure 6.22f).

In the final section (25 km onwards) these parameters return to similar levels observed for the first 7 km over land. Once again breaks in the profile indicate a large proportion of spurious altimeter delays or intermittent triggering.

6.6.4.2 Interpretation

Signals from 0 to 7 km are quasi-specular and comparisons made with Figure 5.10 yield a mean 3 dB backscatter half-angle of less than 3°. Despite the tall narrow peaks, however, some mean waveforms have shallower leading edges, resulting from large-scale surface roughness. The value of \( L_e \) of 1.2 dB km\(^{-1} \) at 4 km inland corresponds with a surface roughness prediction using Brown’s model of the order of 3 m amplitude. Aerial photographs from previous field campaigns reveal that there is a high degree of crevassing and undulating serac ice in this coastal region and this would explain the severe attenuation of many signals. However, the specular component of signals is likely to occur by reflection from wet snow and/or large planar ice surfaces between crevasses. Aerial photography indicates that the surface is markedly less broken between 7 and 25 km inland despite large-scale surface perturbations of over 10 m height and several kilometres in wavelength. The profile in Figure 6f indicates that signals are strong in this part of the basin. The rise in \( \sigma_i^0 \) and similarity of the two backscatter coefficients confirms that returns are diffuse, waveforms periodically having durations of over 300 ns and 3 dB half-angles between 5° and 7°.

It is useful at this point to make comparisons between diffuse land ice signals and ocean returns, and the corresponding responses of the calculated waveform parameters. Ocean returns (diffuse) are a useful yardstick since they result from Lambertian or isotropic surface scatter, tending to the 10° situation in Figure 5.10a (Gatley and Peckham, 1983). Diffuse ocean returns characteristically exhibit high values of \( \sigma_p^0 \), \( \sigma_i^0 \) and \( X_l \), with \( X_e \) returning zeros or very low values. By contrast, quasi-specular returns show high values for \( \sigma_p^0 \) and \( X_e \) with low \( \sigma_i^0 \) and \( X_l \). Surface returns at 14 and 16 km show similar characteristics to ocean signals observed between -5 and 0 km, and \( L_e \) and \( X_l \) return similar values, despite reduced backscatter values. High density snow with a refrozen crust or high water content is a known cause of high reflectances, while small-scale surface roughness is the suggested cause of wider polar scattering.

Between 20 and 25 km mean signals return to values similar to those recorded in the initial seven kilometres over the ice cap, the 3 dB backscatter half-angle falling to around 3°. Scattering contributions are limited to highly reflective surfaces within the pulse limited footprint, antenna beam attenuation effectively suppressing weak returns from areas on the
fringe of the beam-limited footprint. These signals are probably from areas where the snow surface is still relatively dry, and no crevasses have been observed in the upper part of Basin 5. High backscatter values for drier snow have previously been explained by the effects of a complex snow stratigraphy. Snow layering, ice lamellae, and ice glands or lenses produce a component of specular reflection with typical reflectivity of 0.1 (Rott et al., 1985), thus explaining strong backscatter limited to near-normal incidence. Snow pit studies and shallow drilling in Basin 5 during spring 1986 show that such density variations are a common feature of the stratigraphy in this area (Dowdeswell; personal communication, 1986).

6.6.5 Variations in waveform parameters: Vestfonna

6.6.5.1 Results

Waveform parameters from a track across Vestfonna and the central, ice-free valley in Nordaustlandet (Fig. 6.1) indicate three main zones (Figure 6.23). The first is for the initial 12 km, and is from non-glacierized land. The two values of \( \sigma^o \) show several distinct peaks at irregular periods, which are accompanied by rises in \( L_e \) of up to 2.0 dB bin\(^{-1} \) from a level of around 13 dB bin\(^{-1} \). Both attenuation coefficients are extremely variable, \( X_e \) registering a maximum of over 50 \( \mu s^{-1} \) at 6.5 km, but when no signal is evident in late gate bins \( X_l \) returns zeros. Throughout this period spurious delays are recorded, breaks in traces a to f in Figure 6.23 indicating these occurrences.

The second zone, comprising three short sections, is between approximately 15 and 20 km, 24 and 33 km, and 39 and 42 km. All have the same characteristics with relatively steady values of \( X_e \), at around 20 \( \mu s^{-1} \). Although backscatter at nadir (\( \sigma^o_p \)) increases by only 1 or 2 dB, \( \sigma^o \) rises instead by several dB, to plateaus of between 5 and 8 dB. \( L_e \) behaves similarly (as at 16 km) by rising by 0.6 dB bin\(^{-1} \) to 1.9 dB bin\(^{-1} \). The final set of short sections last from approximately 20 to 24 km, 33 to 39 km, and from 42 km onwards. \( \sigma^o_p \) falls by 1 or 2 dB in places to 12 dB, while \( \sigma^o_i \) shows a corresponding decrease to a level of 3 dB. Waveform leading edges have markedly reduced gradients, with minima at 23 and 35 km of 12 and 11 dB bin\(^{-1} \) respectively. During these periods both trailing edge coefficients record a large proportion of zeros (Figures 6.23d and 6.23e).

6.6.5.2 Interpretation

Before crossing the ice streams Bodleybreen, Aldousbreen, and Frazerbreen on Vestfonna the altimeter traverses Helvetesflya, an area of non-glacierized terrain with several melt lakes. At 1 km, over Flysjøen (the largest lake) the signal amplitudes are the highest observed. Although \( \sigma^o_p \) only picks out two definite peaks over lakes (at 1 km and 11 km), \( \sigma^o_i \) has additional
increases at 4 and 5 km. The diffuse nature of signals from Flysjøen, and the 3 dB half-angle of 5°, indicate that the lake surface does not act as a plane reflector, despite minimal metre-scale surface roughness indicated by values for $L_e$ of almost 2.0 dB bin$^{-1}$. Analysis of Landsat imagery showed that Flysjøen was still ice-covered on 12 July. It is therefore inferred that lake ice, accompanied by surface snow dunes, is a likely cause of the relatively wide polar scatter diagram and longer duration returns. Similar waveform characteristics are observed at 4, 6.5, and 11 km, but at 6.5 km $X_e$ exceeds 50 µs$^{-1}$. This occurs because mean signals have markedly shorter duration, due to reflection from a more specular surface such as a stream. Otherwise, between the lakes there are a high proportion of spurious delays and rejected waveforms. This, in combination with stretched leading edges, is a response to partially snow covered terrain having metre-scale roughness of the order of 3 m. Such intermittently snow covered land was observed on the Landsat scene for 12 July.

The second distinguishable set of areas are beyond about 12 km, where the altimeter begins to traverse Vestfonna. This is the first of several snow covered and relatively flat ridges with minimal gradient and metre-scale surface roughness (less than 1 m). They return high amplitude signals. Corresponding 3 dB backscatter half-angles are approximately 3°.

The last group of areas are associated with ice streams flowing orthogonal to the al-
timeter flight-track and between the series of ridges mentioned above (Figure 6.23f). They are identifiable both by their distinct surface character and waveform response. Large amplitude (15-25 m) kilometre-scale surface roughness has been identified in previous work by Dowdeswell (unpublished), but the $L_e$ minimum of 12 dB bin$^{-1}$ at 23 km (Figure 6.23c), for instance, also suggests metre-scale roughness of 3 m or more when compared with predictions for leading edge gradient made by Brown's model. Bodleybreen (Figure 6.23f) has surged since 1976 (Dowdeswell, 1986). Its boundaries are well defined on Landsat and aerial photographic images of the region by heavy crevassing truncated by shear zones (Figure 6.5d). The combined roughness and steep lateral margins of Bodleybreen cause severe attenuation of pulses, and loss of the weakest returns. In regions of broken ice the slope distribution is likely to be markedly different than elsewhere and results in 3 dB half-angles of around 5° over parts of Bodleybreen. The other areas of similar waveform characteristics occur from approximately 33 to 39 km over Aldousbreen, and over the beginning of Frazerbreen from 42 km onwards. Neither has been observed to surge, but both are highly crevassed and are flowing faster than the surrounding ridges (Dowdeswell, 1986). This surface roughness has a significant effect upon waveforms, in contrast with the signals resulting from surrounding ridges. Small scale surface roughness is again liable to cause wider scattering and thus the observed longer duration waveforms.

6.6.6 Conclusions

Results indicate that significant variations in returned altimeter signals occur through changes in the terrain type and surface character of Nordaustlandet ice caps. A number of conclusions may be drawn from this section;

1. When wet snow or ice surfaces are encountered Fresnel reflexion coefficients are large and a specular component is observed to dominate the waveforms.

2. Brown's rough surface scattering model appears valuable as an indicator of the effects of varying surface character on waveforms, where surface scattering is known to dominate.

3. Variations in the observed angular scattering suggest that the scattering properties of snow and ice indeed vary across the ice caps. The mean 3 dB backscatter half-angles of 3° over flat, wet snow surfaces, correspond with the experimental results of Fung et al. (1980), and show similarities with the results from flat coastal regions by Ulander (1985).

4. Significant changes in surface reflexion occur in the early stages of snow surface melt, when free water produced by melting is mainly suspended by necks between snow grains. Previous authors such as Suzuki et al. (1983) have noted its importance in influencing the scattering characteristics of the snow surface.
5. Where large-scale surface roughness and crevassing are minimal, as on Bråsvellbreen, differences between the polar scattering properties of wet and relatively dry snow lead to a significant change in the type of altimetric returns.

Ambiguities remain in the understanding of mechanisms influencing scattering from ice mass surfaces, and how these in turn affect the general characteristics of mean altimeter waveforms. Further experimental work is needed at 13.8 GHz and near nadir into the effects of particular surface properties on radar altimeter waveforms. This should enable development of algorithms to extract data on key glaciological parameters from the forthcoming generation of satellite radar altimeters (e.g. ERS-1). The delineation of dry and wet snow, and bare ice zones on large ice masses from altimeter waveform characteristics should in the future provide valuable information for mass balance studies.
CHAPTER 7

CONCLUSIONS AND RECOMMENDATIONS

7.1 Research Objectives

The main objectives of this research were to:

(1) develop a better understanding of microwave radar altimetric measurements in polar regions, with particular attention to snow and ice surfaces.

(2) examine the utility of a radar altimeter for obtaining geophysical data in glacial environments.

(3) to discuss current and potential applications of radar altimetric data.

To fulfill these objectives the following preliminary studies were conducted:

First, the operating principles of radar altimeters are investigated and discussed, and the altimeter used to collect the data analysed in this thesis is introduced. The RAL instrument and data recording hardware are described and the experiment and MIZEX campaign strategy are outlined to give the background to the overall study.

Second, before altimeter data analysis, a number of pre-processing tasks are completed. These are necessary to be able to obtain statistically valid results and to make valid geophysical interpretations of recorded data;

(a) the statistical fluctuations observed in the raw individual data recorded on magnetic tape require that a data averaging scheme is employed. Various schemes are investigated and a technique known as ‘interpolation tracking’ is adopted to derive mean altimeter signals which do not exhibit fading characteristics. This tracking technique enables accurate superposition of individual waveforms before summing and averaging is performed. Tracking procedures are adapted to account for changes in aircraft altitude, variations in aircraft attitude and rapid changes in the shape of individual pulse returns.

(b) corrections are made to recorded delay values to enable accurate measurements of terrain clearance over ice surfaces; and the implications of the motion of the aircraft for altimetric data collection and pre-processing tracking procedures are discussed.

(c) the inertial navigation data recorded on board the aircraft by the ADDAS system are utilised to obtain a spatial fix, or geo-reference points, for each altimeter data record.
This technique enables the exact location of a data point in space when given the time tag (associated with each record). The procedure has the added advantage that remotely sensed image data which overlap areas of altimetric data collection can be used to support observations made from the altimetric data.

(d) Calibration methods are applied to the radar altimetric data in order that the backscatter coefficient ($\sigma^0$) may be calculated directly from received power values. Different parameters are used for operational calibration on June 28 and June 30, to account for drifts in receiver gain, and to incorporate internal and external calibration data. The final technique utilises measurements made during flights made over radar retro-reflectors and internal calibration files spaced at intervals throughout the data. Power measured at any point during the flight may then be used to derive $\sigma^0$ by comparison with the calibrated retro-reflector response. Since internal calibration files are available at frequent intervals throughout the flight, and since external calibrations were performed on both outbound and homewardbound crossings of Andøya, it is suggested that RAL altimeter $\sigma^0$ measurements are both accurate and precise.

Third, a comprehensive study is made of the electromagnetic properties of snow and ice media. The dielectric properties of a medium are critical to Fresnel reflexion from surfaces exposed to incident radiation from an altimeter, and so the dielectric composition of snow and ice media are calculated using a number of theoretical and empirical mixture formulae. The main observations in the context of altimetry are that increases in both density and water content in snow and ice cause a marked increase in the Fresnel reflexion coefficient at the air/medium boundary. In dry snow, the reflexion coefficient ($R$) increases with density, and the introduction of small amounts of liquid or free water into snow grain interstices causes substantial increases in $R$. Snow structure is recognised as an important consideration because disturbed snow is observed to have different dielectric properties from undisturbed snow and thus different reflective properties. Grain size, in contrast, appears to play an insignificant part in modulating the value of $R$ and is considered of secondary importance. Fresnel reflexion from sea ice surfaces is considered separately from terrestrial snow and ice surfaces, because the introduction of brine pockets into ice media changes the dielectric properties considerably. Two main ice categories are recognised for these purposes, namely, first year and multiyear ice. The justification for such a classification lies in the fact that first year ice generally has much higher brine contents than multiyear ice. Variability in the reflexion coefficient for an air/sea ice interface is large and dependent upon a number of factors, namely; salinity, temperature, density, and brine pocket orientation. In general, the higher the salinity or temperature, the greater the magnitude of $R$. First year ice of high salinity has a value of $R$ which tends to 1 at 0°C. Multiyear ice, in contrast has much lower values of $R$ which vary less rapidly with temperature.
Finally, Chapter 4 extends the discussion of electromagnetic reflection from plane boundaries to more realistic surfaces. At 13.81 GHz, snow or ice surfaces with a small-scale roughness component of 3 mm or more are theoretically adjudged to be electromagnetically rough (using the Rayleigh criterion). It would appear, therefore, that under almost all natural circumstances, snow and ice surfaces may be assumed rough. However, experimentally, the term smooth surface is applied in cases when the observed backscatter coefficient ($\sigma^o$) is in close agreement with values calculated for a specular reflector. Scattering patterns demonstrating a rapid decline in $\sigma^o$ with incidence angle from the normal are termed quasi-specular, and such characteristics can be used to identify relatively smooth surfaces. The contrary response to quasi-specularity is known as diffuse scattering and occurs from rough surfaces. In such cases the decline in $\sigma^o$ with incidence angle is less rapid. The physical optics or Kirchoff approach is used to model backscatter response from polar surfaces which are encountered by the RAL altimeter. The model is developed to demonstrate the effects of surface roughness or surface rms slope upon the surface scattering signature of the ocean surface and various snow and ice media. Results illustrate that at near-normal incidence both coherent and non-coherent scattering components exist under most natural circumstances. The magnitude of each component varies directly with the reflection coefficient at normal incidence and the rms slope of the surface. Using examples from the ocean surface (where the reflection coefficient is assumed to be 1), $\sigma^o$ is shown to depend primarily upon the wave conditions. A fully developed sea causes perfect diffuse scatter or an isotropic backscatter response, whereas a glassy calm sea causes a quasi-specular response. Wave conditions are therefore critical to altimeter response when ocean surfaces are being sensed. One exception to these characteristic scattering responses is identified. Bragg resonant wave components of the ocean surface are also shown to be an effective mechanism for producing quasi-specular returns to wide beam altimeters at high sea states. The tilt modulating effect of low frequency, long wavelength components of the wave spectrum, cause high frequency, short wavelength Bragg scattering wavelets to be presented to incident radiation at the right local incidence angle for phase-coherent scattering to occur.

Surface scattering is the only form of scattering in circumstances where the reflection coefficient tends to a value of 1 (as in the ocean surface example). However, with snow and ice surfaces, the reflection coefficient $R$ is less than 0.1 under most conditions. This means that over 90% of incident radiation is transmitted across the surface interface and into the medium. Penetrating radiation allows scattering to take place from inhomogeneities inside a snow or ice volume. Penetration depths are calculated, therefore, for a variety of conditions. Large penetration depths of the order of several metres are possible in dry snow and pure ice when the dielectric losses and extinction coefficient of these media are negligible. Small amounts of free water or brine reduce penetration depths significantly to less than one wavelength (for water
Volume fractions of 3% or more. Volume scatter contributions to the backscatter response of a surface layer are then calculated using a particle cloud analogy for snow. In dry snow, volume scattering is important, but total volume backscatter nonetheless amounts to much less than that occurring from surface scatter. Surface scatter as a result dominates at near-normal incidence. The volume backscatter coefficient contributes a larger proportion to the total backscatter signature with increasing incidence angle. Volume scattering contributions are negligible under wet snow conditions and for high salinity sea ice.

The physical optics scattering model is adapted into a terrain scattering model in order to add volume scattering contributions from layered media of varying dielectric composition. Results of parametric studies show that dry snow exhibits a different scattering response from wet snow or bare ice surfaces (which appear similar at near-normal incidence angles). With well defined layering in a snow medium, a strong component of interface or surface scatter may occur. Planar dielectric interfaces caused by distinct density horizons or ice lamellae or lenses can cause a coherent component of backscatter from beneath the surface, providing the snow is dry. The terrain scattering model is also modified to investigate sea ice scattering signatures. A parametric study is undertaken using values of salinity, density and snow depth measured during ground truth operations in MIZEX. Results are used to simulate scattering characteristics of East Greenland Sea sea ice during the MIZEX experiment. The main conclusions of this study are that at near-normal incidence the volume scatter contribution is negligible and that wet snow layers can effectively mask the ice surface scattering signature of ice floes. Nonetheless, results suggest that ice floes with a relatively dry snow cover give variable signature with changing salinity and ice surface roughness. These results indicate that sea ice can be classified on the basis of salinity and surface roughness during the late-spring period and shows that it may be possible to distinguish between first year and multiyear ice on this basis from future radar altimetry backscatter observations.

With this essential background to scattering and reflexion from polar snow and ice surfaces the RAL altimeter data collected in marginal sea ice and terrestrial ice regions are analysed.
7.2 RAL radar altimetry in polar regions

The radar altimeter instrument is a proven remote sensing tool in terms of its ability to measure terrain clearance with great precision and to a high degree of accuracy. However, its capacity to obtain useful geophysical information has rarely been investigated. In this thesis the 13.81 GHz RAL radar altimeter is used to examine altimeter operation over polar ice and to identify the main areas of its application in the field of glaciology.

Results of analysis of altimetry from sea ice and terrestrial ice surfaces are presented in chapters 5 and 6 and a number of observations and interpretations are made. Algorithms which enable extraction of geophysical data products from RAL altimetric data are proposed, derived, and tested to demonstrate the utility of the radar altimeter as a remote sensing tool in polar regions. The main conclusions drawn from these studies are outlined below;

- Techniques have been developed to pre-process and view RAL radar altimeter waveforms. Mean return pulse waveforms can be produced using the interpolation tracking technique (Ulander, 1987) for further visual or digital analysis. This pre-processing enabled a detailed investigation of altimetric data from the East Greenland Sea marginal ice zone (MIZ) and two Svalbard ice caps.

- Considerable variations in altimeter signal characteristics occur over the MIZ as a result of changes in backscatter response. Quasi-specular signals are caused by glassy, open water surfaces between ice floes, because water surfaces have a reflectance coefficient a factor of 10 or 20 larger than ice floe surfaces. In ice/water mixtures the altimeter is most sensitive to the proportion of open water within its footprint, and in particular to the surface area sensed within the first few Fresnel zones of its footprint. Signal responses demonstrate that the backscatter contrast between large ice floe surfaces and open water is large.

- Mean altimeter signals are used to reconstruct the variation in $\sigma^o$ with incidence angle, in order to characterise the scattering signature at various points along the track. Descriptive parameters are derived which include two values of $\sigma^o$, the rms surface roughness, and the waveform leading edge gradient. These are used to characterise typical waveform shapes occurring over photographed ice conditions. This technique enables a classification of large ice floes on the basis of the backscatter coefficient at normal incidence, the decay of $\sigma^o$ with $\theta$, and the estimates of rms surface roughness. Observations of ice categories are to some degree validated by aerial photography, which shows differences in floe surface snow cover, pressure ridging, and surface roughness. Ground truth observations of ice salinity, snow depth, and surface roughness to a large extent confirm altimeter detection of bare, first year ice; dry, snow covered first year ice; well ridged, rotted first
year ice, or composite floes; and hummocked, dry snow covered multiyear ice floes.

- It is demonstrated that when ice floes are too small to fill the altimeter footprint area, the growth and decay of waveforms is modulated largely by the liquid ocean surface roughness. Theoretical models of ocean surface scattering reproduce the scattering response of open water surfaces within the MIZ accurately indicating that the influence of ice floes upon high frequency wind-generated ripples and capillary waves may be assumed negligible. The only exception to this rule, however, occurs in situations where grease ice is forming on the surface (see Appendix A). In these circumstances the surface fine structure is damped out. In open, ice-free leads within the MIZ the wind speed algorithm of Fedor and Brown (1982) is tested, yielding values ranging between 0 and 5 m s\(^{-1}\). Predicted windspeeds correspond well with windspeed measurements of 3-5 m s\(^{-1}\) made on that day.

- The altimeter is also shown to sense the slope distribution of the ocean surface within the MIZ. Reconstructions of \(\sigma^0\) with \(\theta\) show that the variations in altimeter waveforms in the MIZ may be used to study the interrelated height and slope distributions of the surface. Observations from altimetry collected along transects from open ocean into the MIZ show a continuous reduction in the height and slope distribution of the ocean surface with distance in from the ice edge. These results are consistent with observations made of wave attenuation in marginal ice. Further to these findings, an adapted Seasat significant wave height algorithm is used as an index of metre-scale surface roughness. Results from this approach support the observation of a decay in metre-scale roughness with distance into the MIZ.

- It is considered that, subject to comparisons with \textit{in situ} data, the SWH measurements from altimeters in open oceans are well understood. However, the operation of the SWH algorithms in MIZ regions is subject to further investigation, and awaits comparisons with wave buoy data collected in the MIZ. Once this approach is validated, it opens up a new area for the derivation of wave penetration information in polar pack ice from altimetric data. Such data are pertinent to the study of the mechanics of and the timing of ice break up.

- In employing some of the main observations made from data acquired in the MIZ, two main algorithms are derived and tested which may be used to derive geophysical data products from an altimeter data stream.

(a) An index of waveform distortion is used to detect ice edges and to locate leads and polynyas. It is based upon the backscattering contrast between sea ice and open water, and calculates a scaled ratio of peak power and waveform energy. The algorithm is shown to locate a MIZ boundary and ice band to an accuracy of 200 m.
(b) An algorithm is developed to calculate ice concentration from the energy in return waveforms. Altimeter waveforms indicate a relatively strong inverse relationship between the integral of signal power and the proportion of ice in the altimeter footprint. The accuracy of this algorithm is calculated at 13 % rms error in concentration estimation, and is of equivalent accuracy to ice concentration algorithms presently employed with satellite passive microwave sensor data. The advantage of an altimeter is that the ground resolution is better than other sensors, though the main restriction is the narrowness of the ground swath. This algorithm needs further development in order to reduce or eradicate the influence of the underlying wave spectrum.

(c) Results indicate that peaks in $\sigma^o$ at normal incidence occur in a direct one-to-one relationship with open water and ice floe surfaces. The only restriction placed on employing a threshold detection algorithm is the restricted range of $\sigma^o$. Although in the case of the RAL altimeter the dynamic range is limited, it is suggested that an algorithm based on simple threshold detection may be used with future satellite instruments employing gain control mechanism. Using such a technique it is suggested that it may be possible to calculate ice concentration statistics along a transect traced by the centre of the footprint.

- Analysis of the RAL altimetric data acquired on June 28 over two ice caps on Nordaustlandet, Svalbard, shows that data quality suffers from slope-induced effects and pulse distortion in steeply sloping regions, in an equivalent manner as GEOS 3 and Seasat. Reflection losses over dry snow, and rough or crevassed surfaces, in combination with these effects, cause large proportions of pulse returns to be mis-triggered. As a result, data analysis is restricted to sections of data where errors due to waveform distortion and mis-triggering are minimised.

- Reconstructions of accurate topographic profiles from corrected RAL altimeter delays show that aircraft radar altimeters are able to profile ice sheet terrain with slopes less than $5^\circ$ to a high degree of accuracy. This is possible providing that instabilities of the platform can be corrected for in the data. Comparison of two profiles with previously recorded Scott Polar Research Institute (SPRI) radio echo sounding flightlines shows that the altimeter successfully tracks the ice surface; and enough detail is picked out to enable interpretations of glacier dynamics and response to bed topography.

- Analysis of waveforms during crossings of coastal ice cliffs reveal a typical sequence of changes in pulse shape. An example from Bråsvellbreen illustrates that coastal ice margins may be located to a high degree of accuracy ($\pm100$ m) using this technique. These data may become a valuable source of accurate information of ice sheet or ice shelf ad-
vance and retreat.

- RAL altimeter waveforms from Nordaustlandet ice caps show that terrestrial ice surfaces have considerable variability in scattering characteristics at normal incidence during the late-spring summer transition. Observed 3 dB backscatter half-angles vary between 3° in coastal regions and between 5 and 7° in central parts of the ice caps. Results correspond with the earlier observations of scattering response from altimetry over Greenland by Ulander (1985).

- Calibration of signal power and a quantitative analysis of waveform shape has enabled two backscatter coefficients, the rms surface roughness, waveform leading edge gradient, and two trailing edge attenuation coefficients to be derived. Rms roughness estimates and leading edge gradients indicate metre-scale surface roughness in the range 0-3 m. Field observations of snow surface micro-relief such as sastrugi are suggested as the main cause of distortions in the waveform leading edges. Supporting data confirms that there is a relationship between waveform properties and surface characteristics, and that the largest values of $\sigma^d$ and smallest 3 dB backscatter half-angles occur over either bare ice surfaces or wet snow surfaces. In contrast, returns from the central parts of the ice cap display lower amplitudes and larger 3 dB backscatter half-angles. It is suggested that the transition between signatures occurs in the region of the dry snow line. These findings are consistent with scattering model results in Chapter 4, and the estimated location of the dry snow line fits with previous observations of seasonal position recorded by Dowdeswell (1984). Complications to recorded scattering signatures may occur in places where the dry snowpack has distinct density variations, ice lenses, or ice lamellae. A strong component of specular reflection occurs in such situations from interfaces within the medium, and results in high amplitude signals with narrow 3 dB backscatter half-angles similar to those occurring from bare ice or wet snow surfaces.

- Scattering variations over terrestrial ice surfaces have not previously been investigated from radar altimetry. Modelling studies in Chapter 4 and supporting data both from Landsat images and ground data collection to a large degree validate observations made from the land altimetry data. It is concluded from these findings that altimeters may in future be used to identify the dry snow line, and also to delineate the ablation zones of ice sheets during late spring and early summer. These conclusions, however, await further geophysical verification by scattering experiments at near-normal incidence.

- With improved tracking agility and large dynamic range, it is suggested that altimeter waveforms received from rough topography may be characterised and used to identify crevassed regions or surging ice streams.

- Experience with GEOS-3 and Seasat radar altimeters has shown that radar altimetric
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- With improved tracking agility and large dynamic range, it is suggested that altimeter waveforms received from rough topography may be characterised and used to identify crevassed regions or surging ice streams.

- Experience with GEOS-3 and Seasat radar altimeters has shown that radar altimetric
data may be used to map ice sheets accurately. Here it is shown that airborne altimeters are an equally valuable tool and an important supplement to future satellite altimeters. Their advantage is that they may be used in regions outside the latitudinal limits of polar orbiting satellite instruments, and in certain circumstances may provide accuracy equal to, or better than that available from a satellite altimeter.

7.3 Recommendations for further work

The limitations of the present results, which can also be viewed as recommendations for future work are:

- A modification of the RAL altimeter is required for the instrument to be of full benefit over surfaces with highly variable reflectivity such as those encountered in polar regions. This would entail replacement of the recording unit by one with a greater dynamic range, to allow greater variation in signal strength about the mean level without saturating. Ulander (1985) suggests that an upper limit 7 dB above the mean signal level is adequate: previous recording devices, as for instance that used on Seasat, allowed at least 10 dB above the mean for fluctuations in pulse return amplitude caused by signal fading.

- The delay trigger mechanism used during data collection over sea ice areas is inadequate when large oscillations in aircraft altitude occur. In future, the video trigger mode of operation is far more suitable and would prevent wandering of pulse returns into and out of the range window of the digitizing unit.

- Scattering characteristics of snow and ice surfaces at near-normal incidence must be investigated further if the full benefit of future altimetric data is to be derived. Most experimental work to date has been conducted at off-normal incidence angles, in support of SAR and scatterometer remote sensing observations. Since volume scattering mechanisms are dominant at these incidence angles throughout most of the year, it is difficult to extrapolate to the near-normal incidence surface scattering response, which is relevant to an altimeter. It is important, therefore, that more field observations of ice surface conditions are obtained. At present very little information is available with which to characterise the scattering signature of these surfaces. Future field experiments must establish which, and in what way, different surface characteristics of snow and ice surfaces affect backscattering at near-normal incidence angles. This must be a combined theoretical and experimental effort, and the use of ground truth data is essential.
• Completion of analyses of other MIZEX '84 remote sensing and ground 'truth' datasets is required before the results of scattering models developed in this study can be tested adequately. Multi-incidence scatterometer data will be an important means of validating their applicability.

• Seasonal variations in scattering response must be investigated to enable interpretations to be made from altimetric data at any time during the year.

• Planning is necessary for the commissioning phase of ERS-1 and other forthcoming satellites carrying altimeters. Simultaneous ground truth operations are a necessary part of any attempts to make accurate geophysical calibrations of instrument data or to validate potential data extraction algorithms.

• The usefulness of ice observations made with future altimeters such as ERS-1 will be strongly influenced by the quality and extent of supporting ground based and airborne validation data. It is suggested therefore that the RAL altimeter is equipped for this task, and of value to future roles in this capacity.

• It is recommended that if the RAL altimeter is to be utilised in future for satellite validation, one critical task should be undertaken. The relative contributions of specular and diffuse returns should be quantified, and their relative magnitude noted from pulse waveforms recorded at different altitudes over the same surface area. This has important implications for equating airborne radar altimeter waveforms with satellite waveforms from the same region.
1 Sea ice: its formation and characteristics

Sea ice is spatially and temporally the most variable of ice masses in the Polar regions. Occupying vast areas of the world's oceans (approximately 10% of the Northern Hemisphere, and 13% of the Southern Hemisphere) it modifies their fundamental character accordingly (Squire, 1984). In turn its character and morphology are dependent upon its age and the climatic and oceanographic conditions prevailing during its growth and lifetime.

Ice growth is well documented in the polar literature (Lewis, 1971; Weeks and Ackley, 1982; Robin et al., 1983 Squire, 1984). When surface cooling of normal sea water (salinity > 24.7°/oo) reduces the upper layer to a point where ice formation may proceed, minute spherical crystals of ice form and rapidly grow into thin discs of 1 – 3 mm diameter and 1 – 10 µm thickness (Armstrong et al., 1966). In calm seas these crystals spread laterally to form a suspension. This gives the sea surface a soupy appearance, damping out small wavelets, and is called frazil ice (see Plate A.1). Unless under calm conditions, ocean waves and currents break up or herd the frazil downwind to form either a thick polycrystalline slurry of frazil platelets (in concentrations of between 20% and 40%) known as grease ice (see Plate A.2), or individual roughly circular plates or 'pancakes' of ice (Martin, 1981). Once formed the growth of the ice proceeds rapidly under the control of oceanic and atmospheric conditions. Fracture into floes may occur at any time during the lifetime of the ice, and normally results from wave action or divergent winds.

Young ice (normally less than 30 cm thick) and first year ice are more saline than older sea ice, and normally three different crystallographic layers may be distinguished. Beneath the snow cover the ice surface remains polycrystalline, with typical salinities between 5 and 15 °/oo. Under this layer is a region of vertical or columnar crystal growth (Langhorne, 1986) having significantly lower salinities (4 - 5 °/oo) due to brine expulsion. During growth, the ice forming at the ice/water interface forms a third layer, known as the skeletal layer owing to its low mechanical strength. As expected this is normally highly saline (30 °/oo).

Second-year and multiyear ice which has survived the summer thaw is very different from younger ice. This older ice occurs in the Fram Strait region by being transported south from the Arctic Ocean in the East Greenland Current. Pressure ridges with sails several metres in height can be formed during its growth, due to tensile and compressional stresses in the ice cover. The summer melt cycle also plays a large part in altering the properties of Arctic sea ice by metamorphosis of the ice surface into melt ponds and hummocks.
Plate A.1  Skim of frazil platelets in suspension, giving the sea a soupy appearance and damping out capillary and small gravity waves.

Plate A.2  Formation of a thick polycrystalline slurry of frazil platelets in a lead between heavily ridged and hummocked sea ice. The photograph illustrates how waves herd the frazil platelets downwind (ie. from left to right) to form 'grease ice', and how effectively waves are damped out in the ice slurry.
Beneath, bulk properties of the ice are completely different, with considerably lower salinities and larger crystal sizes.
1 Antenna beam attenuation pattern corrections

The RAL altimeter beam pattern is critical in the analysis of recorded waveforms, and in particular for the correction of values of signal power recorded in the trailing edges of waveforms. During data analysis, and calculation of the calibrated value of $\sigma^o$, corrections are necessary to adjust for the continuous reduction in gain with increasing angle from the antenna boresight axis. The horn antennas used for pulse transmission and reception have an effective 3 dB full-beam width $\phi_w$ of 10.5° (see Chapter 2 and equation 2.1). The antenna gain $G$ for the ‘effective’ circular beam is approximately Gaussian as a function of the angle $\theta$ from the centre of the beam:

$$G(\theta) = \exp \left[ \frac{-\ln 4 (1 - \cos \theta)}{\sin^2 (\phi_w/2) \cos \theta} \right].$$

Figure B.1 Normalised antenna gain as a function of angle $\theta$ from the antenna boresight axis (ie. angle off-nadir). The 3 dB half-beam width and 3 dB full-beam width are illustrated.
Figure B.1 illustrates the antenna beam attenuation pattern, normalised to the value of maximum gain along the centre of the beam (ie. antenna boresight axis). The effective 3 dB half-beam limit of the horn antenna is nominally 5.25°, and represents the point at which received power is reduced to half of its maximum value.

In practice, to be able to calculate the variation of $\sigma^\circ$ with $\theta$, one must assume that the peak power recorded in a waveform occurs from normal incidence backscatter at nadir. A power value recorded at a relatively later delay, in a subsequent range bin $n$, represents backscatter from a range ring annulus which is displaced from nadir by the distance $\sqrt{H \cdot c \cdot \tau_n}$, where $H$ is the aircraft height, $c$ is the velocity of light, and $\tau_n$ is the delay relative to the time of receipt of peak signal amplitude. $\tau_n$ may be calculated from simple geometry by

$$\tau_n = \frac{H(1 - \cos \theta)}{c \cos \theta}.$$  \hspace{1cm} B.2

![Figure B.2](image_url)

**Figure B.2** Normalised gain reduction with varying delay displacement $\tau_n$. The maximum gain of the antenna is 23.3 dB along the boresight axis.

A correction factor $\delta$ is introduced to correct power values in the trailing edges of waveforms to be used in further analysis, or in calculation of a calibrated coefficient of backscatter.
The corrected power value \( P(n) \) in range bin \( n \), is

\[
P(n) = A \exp[-\delta \tau_n]
\]

where \( A \) is the power value at peak amplitude (ie. \( n = 1 \)), and \( \delta \) is the antenna beam attenuation correction factor, where;

\[
\delta = \frac{\ln 4}{\sin^2(\phi_w/2)} \frac{c}{H}
\]

Figure B.2 Shows the normalised gain reduction with bin displacement from waveform peak amplitude, and illustrates the attenuation of power in in range bins in waveform trailing edges. Backscatter is attenuated by 3 dB within relative bin displacements of only 40 bins. Thus, the greater the value of \( \tau_n \), the larger the correction factor.
REFERENCES


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Vant, M.R., 1976. A combined empirical and theoretical study of the dielectric properties of sea ice over the frequency range 100 MHz to 40 GHz. [Ph.D. Thesis, Carleton University, Ottawa, Ontario, Canada, 1976.]


**ADDENDUM TO REFERENCES**

