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OBSERVATIONS OF THE SURFACE PROPERTIES OF THE ICE SHEETS BY SATELLITE RADAR ALTIMETRY

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Abstract. By comparing modelled and averaged satellite altimeter return, it is demonstrated that time profiles of altimeter return can be used to provide important information on the surface properties of the ice sheets. Altimeter ice-sheet radar echoes from low altitudes and/or relatively low latitudes are, in general, dominated by surface scattering and, in Greenland, the area of surface-dominated return broadly coincides with the zone of summer melting. Seasonal variations in the echo wave-form shapes are negligible in all regions studied, with the possible exception of an area near the margin of the Greenland dry-snow zone. In general, the model explains well the observed variations in mean wave-form shape, but small discrepances between the model wave forms and the recorded wave forms indicate that sub-surface layers may be influencing the shape of the return. The possibility of deriving quantitative estimates of surface properties is explored by fitting model returns to averaged altimeter wave forms from the Wilkes Land plateau in Antarctica. Surface roughness can be measured unambiguously from the wave-form data, but estimations of other parameters, such as grain-size, snow density, and snow temperature are found to be ambiguous because different surface parameters have a similar influence on the shape of the return. Despite this, the derived estimates compare well with ground-based observations and suggest that the satellite altimeter may have an important role to play in providing information on the surface properties of the ice sheets.

INTRODUCTION

The satellite altimeter is designed to measure the range of the surface from the satellite, to record a part of the received echo, and to estimate the magnitude of the return power. The satellite range has been used to obtain measurements of surface topography over the ice sheets with a precision of the order of tens of centimetres (Zwally and others, 1983), and this is an order of magnitude better than the precisions obtainable by most conventional survey techniques. Although the measurement of surface elevation is the most well established application for ice-sheet altimetry, some recent papers have highlighted less obvious applications, mostly in connection with the monitoring of ice-shelf features (Partington and others, 1987; Zwally and others, 1987; Ridley and others, in press). These papers show that important information concerning the characteristics of the surface can be derived from the shape of the return and recent work has begun to provide an indication of how, in more general terms, surface properties influence the shape of the altimeter return (Ridley and Partington, 1988). In particular, it has been demonstrated that, over much of the ice-sheet plateau, the altimeter return can be influenced by volume scattering and can thus be sensitive to such parameters as grain-size, snow-packing properties, and snow density.

The aim of this paper is to extend this work by investigating spatial and temporal variations in ice-sheet altimeter returns and interpreting the observed variations using the model developed by Ridley and Partington (1988). The aim is to demonstrate that the altimeter can be used to investigate spatial variations in such parameters as grain-size and surface roughness, providing a useful tool for workers interested in surface accumulation and ablation, and the distribution of ice types. The results will also have implications for the interpretation of synthetic aperture radar data, the measurements of ice-sheet surface elevations, and the long-term monitoring of ice-sheet mass balance by altimetry.

A brief review of the model developed by Ridley and Partington (1988) is provided, followed by a description of the data and processing techniques. The second half of the paper describes a comparison of the model with the data and draws implications regarding the measurement of geophysical parameters using the satellite altimeter.

THE MODEL OF RADAR-ALTIMETER RETURN FROM ICE SHEETS

In general, the radar echo returned from an ice sheet consists of both surface- and volume-scattered components. Ridley and Partington (1988) have developed a model of the return, based on certain simplifying assumptions concerning the nature of the ice-sheet surface. This model is reviewed here; surface scattering is considered first, followed by volume scattering.

The model used by Ridley and Partington (1988) to describe the surface-scattered component of the return is similar to that used by Brown (1977) for ocean return, and is based on physical optics theory. The instantaneous surface-scattered return is assumed to be determined by the surface area simultaneously scattering radiation and by the shape of the antenna pattern. For the interpretation of Seasat and Geosat data, the latter is assumed to be Gaussian with a 3 dB beam width of 0.8°. The surface is taken as planar and horizontal, and the reflectivity of the surface is taken to be constant over the beam width of the antenna. Surface roughness is modelled under the assumption that there is a very large number of scattering facets, that their height distribution follows a particular functional form, and that facets located at any particular surface height are distributed randomly within the footprint. This latter assumption is equivalent to assuming that the dominant surface-roughness wavelengths are sufficiently small and

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randomly phased, and amplitudes sufficiently large (to first order larger than the radiation wavelength), that the back-scatter coefficient does not change significantly through the incidence angles sampled by the altimeter (0–~0.4°). With these assumptions, it is then valid to generate a wave form numerically by aligning and then summing a large number of flat surface returns so that the delay-time distribution of the flat-surface wave forms corresponds to the required height distribution of scattering facets. This enables different height-distribution functions to be incorporated easily into the model. The particular height-distribution function used here, however, is the Gaussian. Note that this model of surface roughness does not allow for any dominant surface-roughness wavelength.

In contrast to previous models, the Ridley and Partington model accounts for both surface and volume scattering, although under the following simplifying assumptions:

The material properties are statistically homogeneous across the footprint, and are constant, or vary linearly, with depth.

The snow consists of a mixture of dry, spherical ice particles and air.

There is no multiple scattering.

The volume-scattered part of the model return is generated by considering both the dielectric properties of the snow and the geometry of the pulse interaction with the surface. Mixing formulae are used to relate snow temperature, density, Formzahl, and grain-size to the real and imaginary parts of the dielectric constant of the snow. Formzahl is a parameter used to express the packing properties of the grains, and has been defined by Wiener (1910). The dielectric constant can then be used to derive values for the scattering and absorption coefficients, the former under assumptions of Rayleigh scattering. The model volume-scattered echo is then generated numerically by considering the geometry of the pulse interaction with the surface, incorporating surface roughness in the same way as for surface scattering. The Ridley and Partington model, modified to account for the different propagation speeds of radiation in snow and air, is as follows:

$$E_{\mathrm{r}} = M \int_{t_{\mathrm{n}} - \tau/2}^{t_{\mathrm{n}} + \tau/2} \left\{ \frac{1 - \exp\left[2\alpha c(t_0 - t)/\eta\right]}{\left[ct/\eta + ct_0(1 - 1/\eta)\right]^2} \right\} \mathrm{d}t\,. \tag{1}$$

Here, E_{r} is the received power in a single-range bin, characterized by duration τ, c is the propagation speed of the radiation through a vacuum, η is the refractive index of snow, α is the extinction coefficient, t_0 is the time at which the pulse intersects the surface, and M is defined as follows:

$$M = \frac{6E_{\mathrm{t}}T^2\lambda^2G^2\left[1 - \mathrm{e}^{-\alpha_{\mathrm{s}}\mathrm{d}r}\right]}{512\pi} \tag{2}$$

where E_{t} is the transmitted power, T is the surface transmission coefficient, λ is the radiation wavelength, G is the antenna gain, α_{s} is the scattering coefficient, and r is the satellite range.

Figure 1 shows the effect on the shape of the return of various surface parameters. It can be seen that volume scattering distorts the leading edge of the echo-time profile, creating a "rounded" plateau region. Grain-size and snow density influence the rapidity with which the magnitude of the return reaches a constant value after initial pulse intersection with the surface. Although temperature influences the shape of the return in a similar manner, the effect is very small for temperatures well below 0°C. Surface roughness, which is given in Figure 1 as a root-mean-square value for the Gaussian distribution (in m), has a most marked influence on the early part of the return. If the echo is dominated by volume scattering,

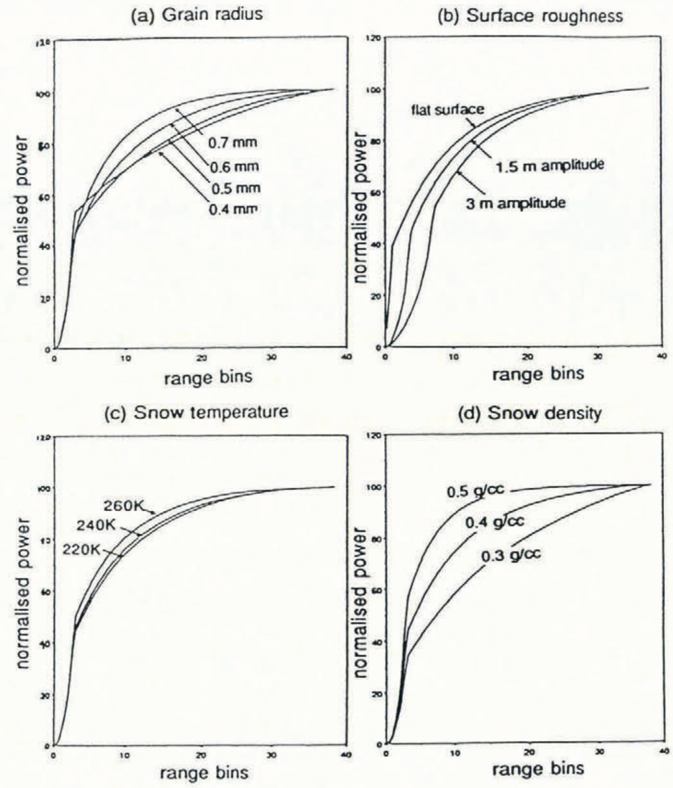

Fig. 1. Result of varying snow parameters on the shape of the radar return. The horizontal axes represent range bins and the vertical axes represent normalized signal amplitude (from Ridley and Partington, 1988). Note the similar effects of grain radius, snow density, and temperature on the shape of the return and the unique effect of surface roughness.

surface roughness creates a "toe" at the foot of the leading edge. If the echo is dominated by surface scattering, the entire leading edge is determined by the surface roughness, as for ocean scattering (Brown, 1977). The results of this modelling suggest, importantly, that it is possible to estimate ice-sheet surface roughness using the altimeter return, but that there are ambiguities involved in distinguishing between grain radius and snow density.

Over a diffuse surface, in which the addition of radiation phase against time is random, the differential of an altimeter-echo wave form corresponds to the height distribution of surface scatterers. The differential, $f'(n)$, is defined as follows:

$$f'(n) = \frac{\delta(R(n))}{\delta(n)}. \tag{3}$$

Here, n refers to the range and $R(n)$ refers to the amplitude of the return at that range. Over the ocean, $f'(n)$ is Gaussian to first order. If volume scattering is important, the differential is skewed, the degree of skewness being a measure of the depth of penetration. Figure 2 shows, in simplified terms, differentiated return expected from surface scattering and volume scattering. Note that, in the case of significant volume scattering, the differential no longer provides an indication of the height distribution of surface scatterers. The skewness of the differential provides a convenient means of assessing the relative importance of volume scattering.

In summary, the model can be used to provide an indication of the scattering properties of the surface and thus an indication of the physical properties of the surface, although the model results do suggest that there are ambiguities involved in distinguishing between the effects on the shape of the return of most of the physical properties of the surface, except in the case of surface roughness.

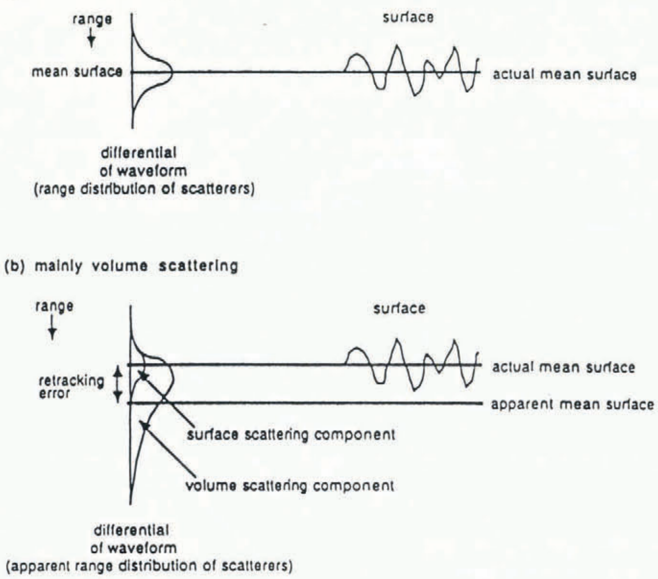

Fig. 2. Diagram showing simplified differentiated mean wave forms from regions generating surface scattering (a) and mainly volume scattering (b). The first is nearly symmetric and the second is positively skewed.

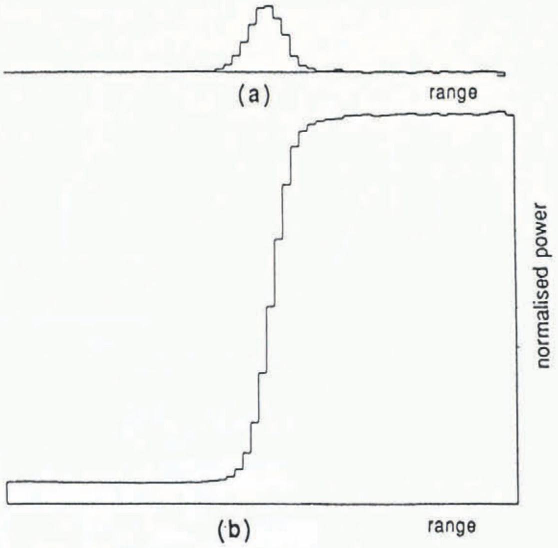

Fig. 3. Averaged ocean altimeter return (b) with differentiated return (a) generated from 1000 telemetered returns from the South Atlantic Ocean.

DATA PROCESSING

Data processing is required to ensure that the assumptions of the model are met as nearly as possible in the data being analysed. Most individual ($10\,s^{-1}$) wave forms recorded by Seasat and Geosat do not conform to the model shape. This is probably a result of variations in surface dielectric and geometric properties within the footprint of the altimeter creating a form of noise in the $10\,s^{-1}$ wave forms which can be averaged out to leave a mean surface-height distribution and mean volume-scattered return. An averaging technique has been used to increase the sample footprint area, thereby generating greater statistical homogeneity and enabling the model wave forms to be compared with, and fitted to, the data. Summation of wave forms is not trivial because the position of the wave form within the range (sample) window varies in response to tracker and surface-range movements. To average the data correctly, a centre of gravity tracker developed by Wingham and others (1986) is used. The method effectively approximates the wave form by a rectangle and aligns wave forms for summing at the position of the leading edge of the rectangle: the rectangle is chosen to have a centre of gravity and area equal to that of the wave form. The position alignment, p, measured in range bins from the beginning of the range window, is taken as:

$$p = \underbrace{\frac{\sum_{n=1}^{n=60} nR^2(n)}{\sum_{n=1}^{n=60} R^2(n)}}_{\text{A}} - \underbrace{\frac{\left\{\sum_{n=1}^{n=60} R^2(n)\right\}^2}{2\sum_{n=1}^{n=60} R^4(n)}}_{\text{B}} \qquad (4)$$

where $R(n)$ is the amplitude of range bin n ($1 < n < 60$) and $R^2(n)$ is the square of the amplitude of range bin n. The first part of the equation (A) gives the centre of gravity of the wave form and the second part (B) gives half of the width of the rectangle. Thus, (A − B) gives the position of the leading edge of the rectangle. This retracking technique uses all range bins to estimate the retracking position, and thus is a considerable improvement over threshold retracking techniques.

The averaging procedure can be tested by generating a mean ocean return, which is shown in Figure 3. The lower plot shows the average of 1000 wave froms recorded over the southern Atlantic Ocean. The upper plot shows the differential of the wave form, and this is shown in all plots to facilitate the distinguishing of surface- and volume-dominated wave forms. The wave forms have been corrected for antenna and range window gain, and because the reflectivity of the ocean surface varies little across the footprint, the differential provides the probability density function (pdf) of the surface facets with range. The averaging procedure can be validated by comparing the differential with the expected height distribution. The range pdf of the ocean surface is almost Gaussian, but skewed slightly as a result of electro-magnetic and surface-facet distribution bias towards the surface troughs (Srokosz, 1986): normalized skewness (β_1) and kurtosis (β_2) are 0.116 and 3.321, respectively. The function is, therefore, close to normal ($\beta_1 = 0$; $\beta_2 = 3$) and so the averaging technique can be assumed to introduce negligible bias into the shape of the wave forms.

Before averaging, data corrupted by various intrumental artefacts are discarded. Over the periphery of the ice sheet, the tracker failed to maintain lock and, even where data were available, the wave-form leading edges were often attenuated, indicating large surface slopes. In these areas, the assumption of a horizontal surface is clearly invalid, and so analysis was restricted to the ice-sheet plateau (slopes <0.5°) and ice shelves.

The specific areas chosen to provide data for averaging are shown in Figure 4, together with the locations of the eight wave-form averages generated by Ridley and Partington (1988). These areas cover a wide range of altitudes and latitudes. Averaging was carried out over two different sizes of region. Large regions were used to investigate broad variations in wave-form shape over the scale of the ice sheets themselves and over the time-scale of a year. Areas chosen ranged in size from about 23 000 km² (Wilkes Land) to 67 000 km² (Greenland: lat. 70.5°N.) and included the ice shelves labelled in Figure 4, parts of Greenland, and parts of Wilkes and Enderby Lands (the latter divided into 20° longitude "strips" at lat. 72°S., where ground-track spacing is at its most dense). In each case, a thousand telemetered wave forms were averaged. Smaller regions were used for fitting of the averaged wave forms by the model ice-sheet altimeter return, to enable investigation of changes in surface properties over the scale of tens to hundreds of kilometres. One thousand wave forms were averaged from each of 16 smaller regions in Wilkes Land (5° longitude "strips" at lat. 72°S.). The regions were still sufficiently large to enable statistical noise in the wave forms to be averaged out. The regions chosen were at lat. 72°S. to provide a high density of data points and the areas chosen were 34 km by 170 km in size. As the regions are located on the ice-sheet plateau, it was considered likely

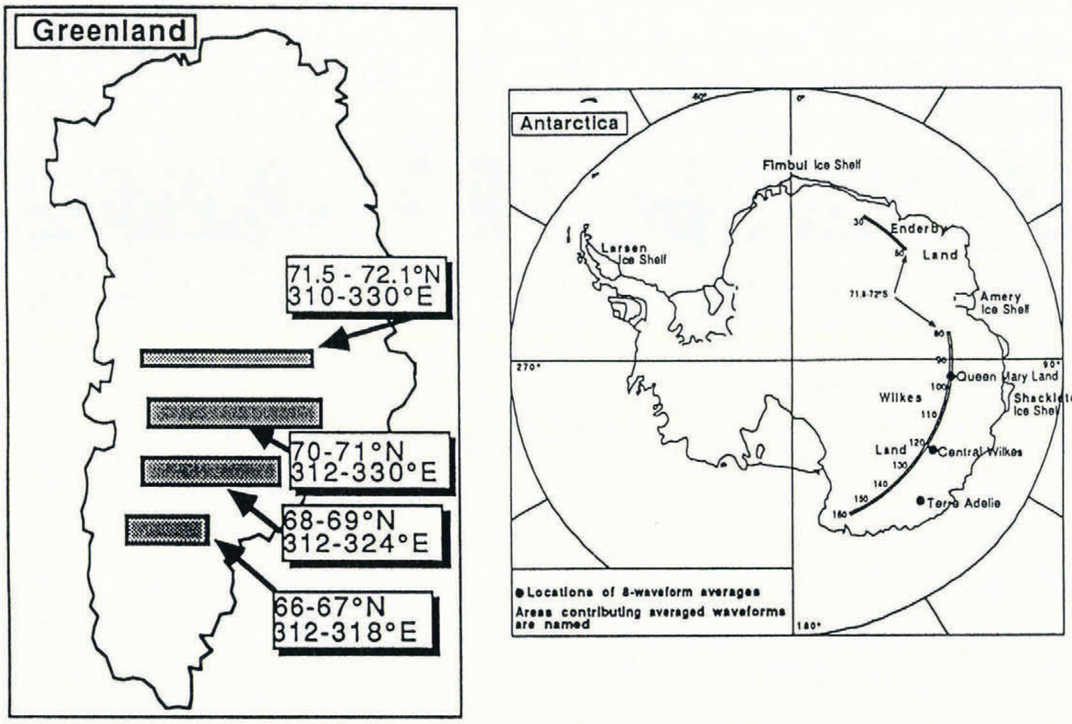

Fig. 4. Locations of regions contributing averaged wave forms in Antarctica and Greenland, including the regions contributing the eight wave-form averages discussed by Ridley and Partington (1988).

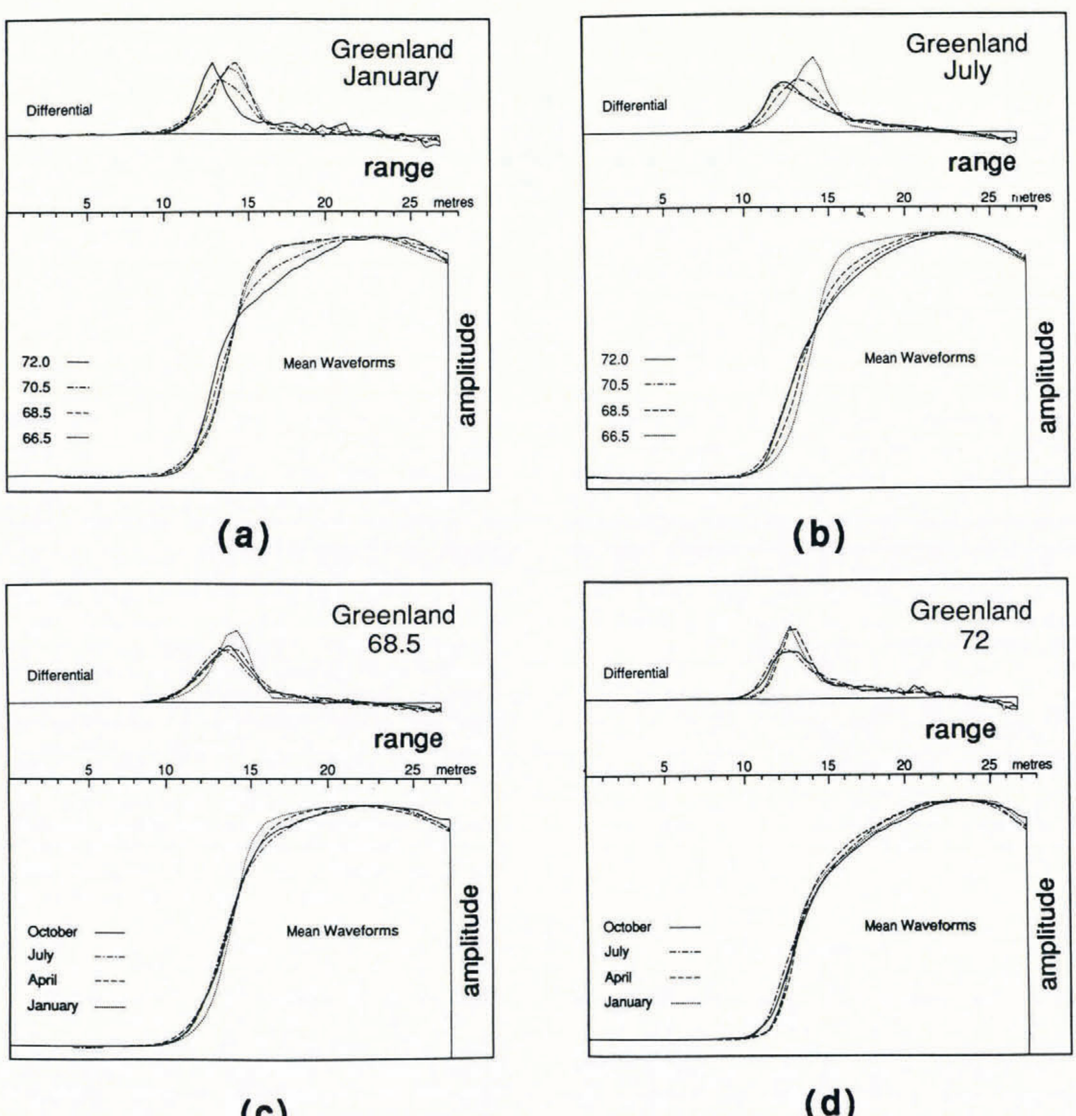

Fig. 5. Averaged Greenland ice-sheet return and differentiated mean return (the former normalized in the vertical direction). Averaged wave forms are all normalized in the vertical direction. Note the increased influence of volume scattering towards the north of the ice sheet (a) and (b).

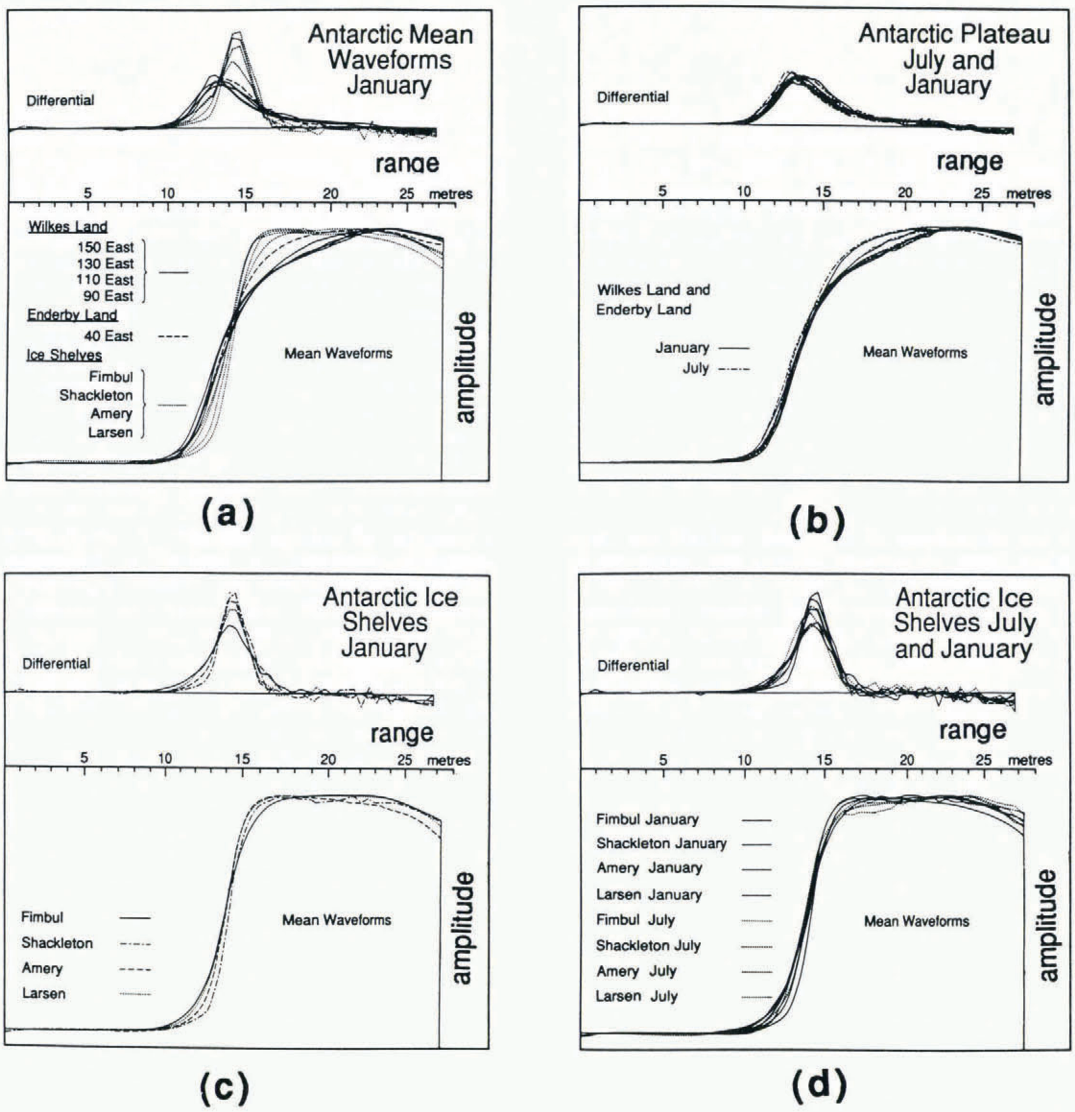

Fig. 6. Averaged Antarctic ice-sheet return and differentiated mean return (the former normalized in the vertical direction). The Wilkes Land wave forms are from 20° longitude parts of the lat. 72°S. region shown in Figure 3, and the Enderby Land wave form is from long. 30°–50°E. Centre longitudes are used to distinguish regions at lat. 72°S. Particularly worthy of note are the marked differences in wave-form shapes recorded over the ice shelves and the high-altitude ice sheet (a) and the small differences in wave-form shapes recorded from different ice shelves (c).

that surface conditions were homogeneous over such an area. In practice, there is little information which can be used to verify this, other than the quality of the data fits by the model and so the size of the areas was somewhat arbitrary. The fitting of the model wave forms to the data was carried out by least-squares methods.

To compare data on a seasonal, as well as a regional, basis, the data used to produce an averaged wave form were recorded within a period of 14 d: data from a shorter period would have provided inadequate coverage over some areas. Geosat data from the *last* 14 d of April, July, and October 1985, and the *first* 14 d of January 1986, were each averaged to provide time series of mean wave forms.

RESULTS

Figures 5 and 6 show the averaged altimeter wave forms from the larger regions of the ice sheets. Plots of the averaged wave forms show that wave-form shape varies greatly with location on the ice sheet, but that seasonal variation is much less significant. Some averaged wave forms are close in shape to that of return from the ocean (Fig. 3). These include all the ice-shelf wave forms and Greenland wave forms south of about lat. 68°N. These will be referred to as *surface-dominated* wave forms. Elsewhere, wave forms with asymmetric leading edges are recorded. In the case where volume-scattering appears dominant, these will be referred to as *volume-dominated* wave forms. Intermediate wave forms exhibiting both surface and volume scattering will be referred to as *transitional*.

This interpretation of the averaged altimeter wave forms is justified on several counts. First, for the so-called volume-dominated wave forms to be the result of surface-scattering, the surface-height distribution indicated by the differentials would have to represent r.m.s. surface roughness of the order of 10 m on the ice-sheet plateau. Over the sample footprint of a single altimeter wave form, which is about 10 km in diameter, this is in general unlikely. Secondly, wave forms in Figures 5 and 6 are consistent with the model results. Thirdly, estimates of dielectric properties of the ice-sheet surface calculated using measured snow properties suggest that volume scattering should, in general, dominate the return unless the surface has been subject to crust formation, and these estimates are supported by experimental evidence, some of which have been reviewed by Ridley and Partington (1988). Note that the curves in Figure 1 show discontinuities (or inflection points) associated with the position at which the back of the pulse shell intersects the surface. It is important to note that such discontinuities will not be present in the data analysed here, because of the data-averaging procedure.

Ice-shelf wave forms

Averaged wave forms from the Antarctic ice shelves are shown in Figure 6. All the wave forms are

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surface-dominated and the variations in shape across the ice shelves and at two different times during the year are very small. The domination of ice-shelf return by surface scattering may be due to the presence of a near-surface crusty layer produced by summer freeze-thaw cycles, wind, or radiation. Doake (personal communication) reports that such features exist on the Larsen Ice Shelf and are rarely more than 0.5 m below the surface, even during winter, which may explain why there is so little variation in shape with season. As the wave forms are surface-dominated, surface geometry is effectively the only factor which will influence the shape of the return. Ridley and Partington (1988) demonstrated that, to first order, the assumption of Gaussian surface-height distribution, as encapsulated in Brown's (1977) model, is realistic for data from the Larsen Ice Shelf, and the length of the leading edge can be used to locate areas of large surface roughness, as in the case of ocean return (Partington and others, 1987; Ridley and others, in press).

Greenland wave forms

Returns from the Greenland ice sheet include all three categories of wave form. A sample of averaged wave forms taken along the Greenland plateau from lat. 72° to 64°N. shows a clear relationship between the relative importance of surface and volume scattering, and latitude (Fig. 5). Returns received from north of about lat. 68°N. exhibit significant surface penetration, indicated by the wave-form shape. The areas from which these wave forms were recorded coincide partially with the dry-snow zone mapped by Benson (1962). Snow which does not melt is less likely to form a hard surface crust (although crusts can also be formed by wind and radiation). With no surface crust, the high surface-transmision coefficient of snow will result in domination of the return by volume scattering.

The relationship between diagenetic facies and the relative contributions of surface and volume scattering is shown in Figure 7. Returns from south of lat. 68°N. exhibit little sign of penetration and coincide fully with the percolation zone. Summer melting is likely to form a crust and reduce the surface-transmission coefficient. An interesting feature of the Greenland data is the seasonal variation in the shape of the wave forms at lat 68.5°N. The data suggest that there is more surface penetration during July than January. It might be expected that any observed changes would indicate less penetration during the summer, because of increases in temperatures. There are a number of possibilities. First, the distribution of altimeter ground tracks within the sample areas may have to be changed. As the region lies at the edge of the dry-snow zone, it may be expected that the apparent ratio of surface-to-volume scattering would be sensitive to the precise distribution of altimeter tracks across the region. Alternatively, there may be real changes in surface-scattering properties.

If these changes in wave-form shape are caused by real differences in the scattering properties, it will be necessary to account for them when monitoring surface elevation, otherwise erroneous elevation changes of tens of centimetres or more will be measured, as described by Ridley and Partington (1988). These results emphasize the importance of taking account of wave-form shape when retracking. However, a method of retracking 10 Hz wave forms which accounts for volume scattering has not been developed as 10 Hz wave forms are strongly influenced by local topography and, possibly, vertical variations in dielectric properties within the altimeter footprint. This is an area in which further research is required if topographic changes of the order of tens of centimetres are to be detected by satellite altimeter.

On a more positive note, these results suggest that it may be possible to locate the edge of the percolation zone using altimeter data, although it is not necessarily easy to distinguish between areas of dry surface crusts and areas of partial surface melting. The precision with which such zones can be delimited is not easy to determine. The along-track measurement spacing with ERS-1 will be of the order of 330 m, and the spacing of ground tracks, given a 35 d repeat cycle and a latitudinal limit of ±82°, would vary

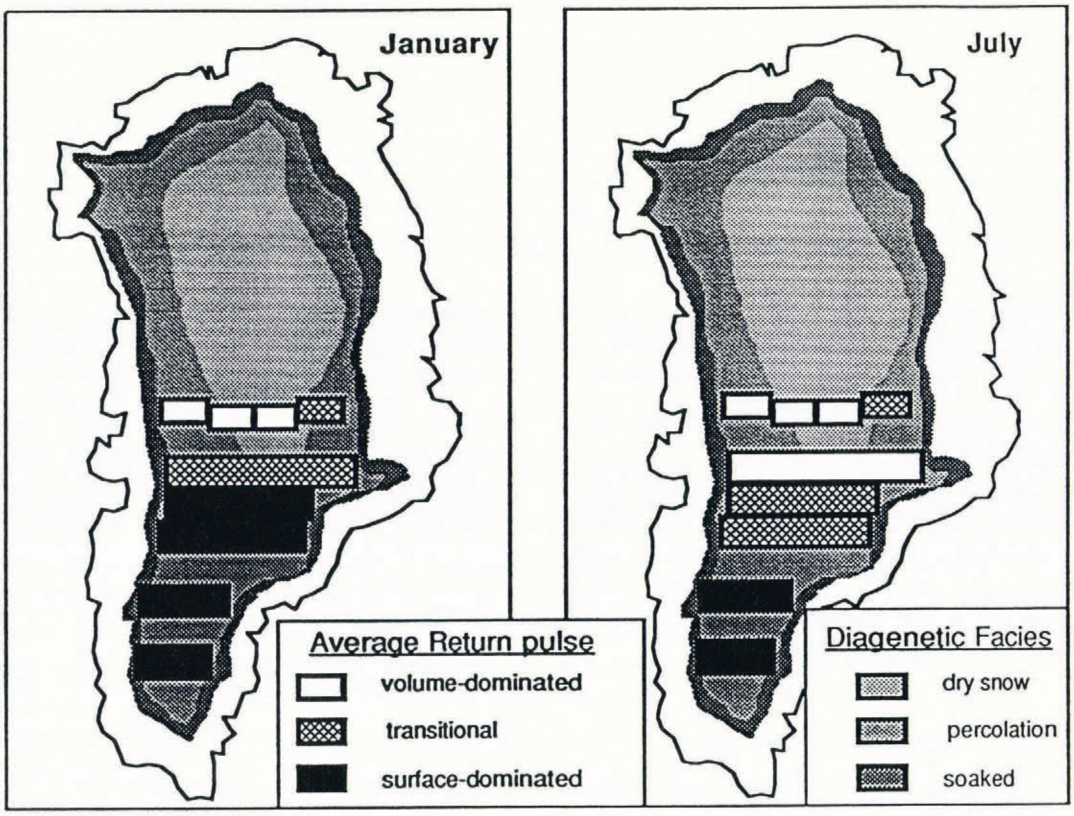

Fig. 7. Classification of mean Greenland wave forms (for January and July) superimposed on a map of diagenetic regions reproduced from Benson (1962). The soaked snow zone is defined as that region where the entire previous winter's snowfall is raised to melting point by the end of the summer. The percolation zone is the region of surface melting during summer (snow at depth may remain below 0°C). The dry-snow zone is defined as that region where there is normally no surface melting, even in summer. Note the correlation between the diagenetic facies and the proportion of surface-to-volume scattered return.

from about 15 km in southern Greenland to 0 km at the latitudinal limit of 82°N. However, the wave forms analysed here are averages and it is not clear whether individual wave forms can yield unambiguous information on the degree of surface penetration. It may be necessary to generate wave forms which are running averages along the satellite track before the model can be used to locate different diaganetic zones, in which case the along-track resolution may be worse than 330 m. However, an along-track resolution of 3 km would provide far better resolution than currently exists and, if demonstrated, the technique could be used to monitor the extent of the percolation zone and possibly to improve mass-balance estimates.

Antarctic plateau wave forms

Figure 6 shows averaged wave forms from Wilkes and Enderby Lands. The mean wave form from Enderby Land is clearly transitional in form between those from the ice shelves and Wilkes Land. The sampled region of Enderby Land is at markedly lower altitude than that of Wilkes Land, suggesting that, in addition to latitude, the shape of the return is related to altitude. The degree of radar penetration may also be related to surface slopes, as these determine the strength of gravity-driven katabatic winds, and to accumulation rates which, in general, decrease inland. Without additional data, this must remain speculative. However, Fujii and others (1987) confirmed that there is glazing on the surface in Enderby Land, which they claimed is persistent and which may account for reduced penetration in this region. There is very little, if any, seasonal variation in the shape of the return.

The regional differences in wave forms are therefore consistent with *a priori* expectations. Large-scale variations in the shape of the wave forms appear to be related to altitude and latitude which are presumably acting as proxies for physical conditions such as mean annual temperature, mean maximum temperature, precipitation, and mean wind intensity. In general terms, volume-dominated wave forms tend to be recorded in dry-snow regions, where there is no surface melting during summer. However, it is possible that surface-dominated wave forms are recorded not only in areas of summer melting but also in areas of strong katabatic winds, where crusts form, and in areas of "blue ice". Clearly, there is scope for continued systematic research into types of surface snow which are found and the conditions under which they develop. Such studies may enable the interpretation of altimeter data to be improved.

Having demonstrated that the model is useful as a qualitative tool for the interpretation of altimeter data, the potential of the model as a quantitative tool is investigated. Estimates of surface parameters are obtained by fitting the model to the averaged data, and the reliability of the model can be assessed by comparing the estimates with ground-based measurements. To this end, averaged wave forms from relatively small areas of Wilkes Land have been fitted using the model of Ridley and Partington (1988) and some of these are shown in Figure 8. The parameter values measured by fitting the model to the wave forms are shown in Table I.

The parameter values obtained from the fitted model are consistent with reported observations of such parameters (e.g. Kuroiwa, 1956; Bender and Gow, 1961; Gow, 1971; Alley and others, 1982) and measured dielectric properties (e.g. Nyfors, 1982), and a brief summary of some of these reported observations has been provided by Ridley and Partington (1988). However, it should be noted that the temperature adopted is somewhat arbitrary because the model is insensitive to changes in temperature of as much as 20°C, for temperatures well below 0°C (Fig. 1) and, in addition, the effects of grain-size and snow density on the return are similar, so there is some ambiguity involved in measuring these two parameters. However, measurements of surface roughness and κ are not ambiguous, as these parameters have a unique influence on the shape of the return, and so important geophysical measurements may be derived using this technique. The precision with which surface roughness may be measured using this technique is the order of 10 cm. Variations in mean surface roughness across the region are quite marked, the mean surface roughness being greatest towards the ends of the region. Surface roughness is likely to be related to the strength of winds. The parameter κ, providing a measure of the relative importance of surface and volume scattering, varies quite considerably, even on the plateau, and appears to be related only partially to altitude. It is likely that κ is related also to the strength of winds and large surface coefficients may be expected in regions of glazed ice, so an imperfect

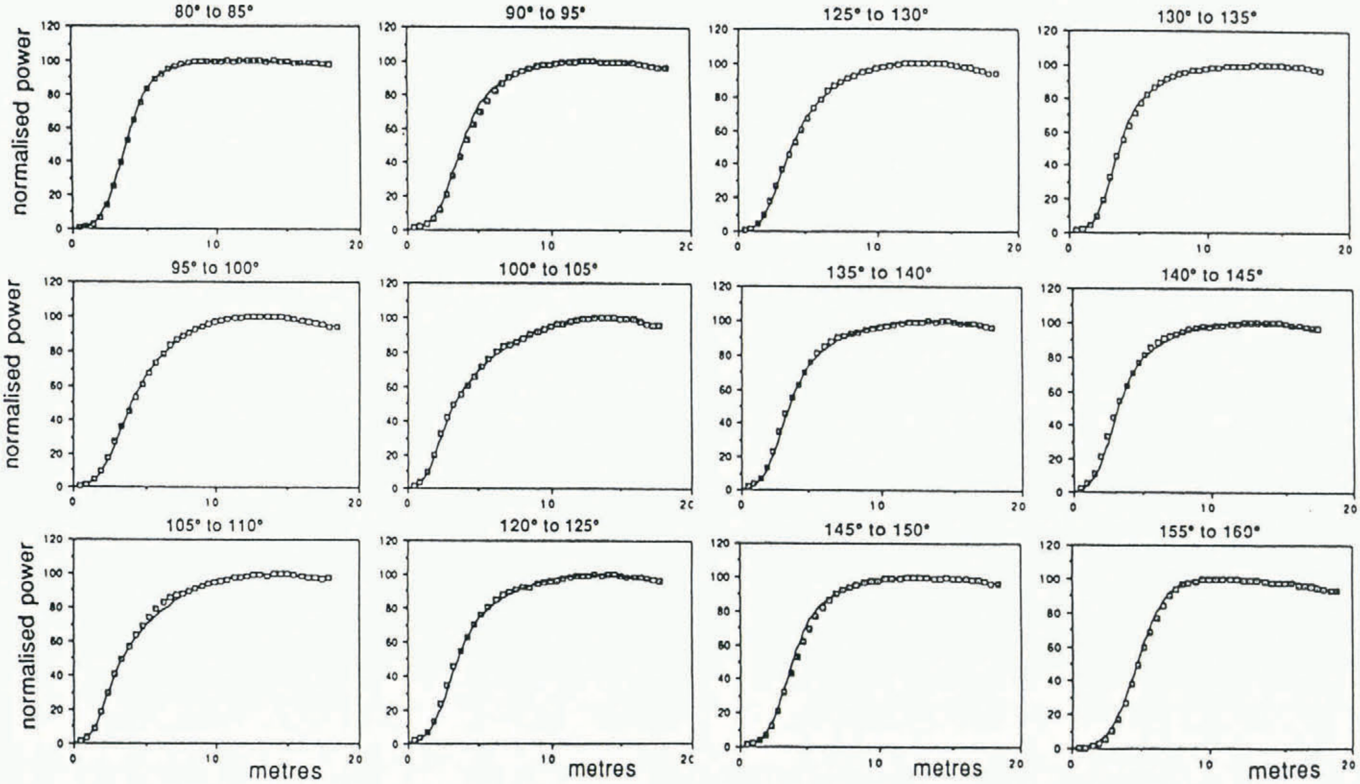

Fig. 8. Averaged wave forms from 5° longitude parts of Wilkes Land (squares), together with model ice-sheet wave-form fits (lines). The units of the horizontal axis are tens of centimetres. The plots are normalized to 100 in the vertical direction.

TABLE I. MODEL PARAMETERS USED TO FIT WILKES LAND WAVE FORMS

Longitude (all 71.80–72.06°S.)	*Altitude*	*Roughness (r.m.s.)*	*Proportion of surface-to-volume scattered return*
°E.	m	m	κ
80–85	2855	1.2	0.5
85–90	N/a	N/a	N/a
90–95	3225	1.3	0.2
95–100	3195	1.0	0.01
100–105	3026	0.7	0.01
105–110	2807	0.7	0.01
110–115	2800	0.7	0.01
115–120	2807	0.7	0.01
120–125	2919	1.0	0.1
125–130	3042	1.0	0.01
130–135	2952	1.0	0.15
135–140	2527	1.1	0.1
140–145	2275	1.0	0.12
145–150	2106	1.2	0.2
150–155	2047	1.6	0.5
155–160	2123	1.6	1.0

Temperature = −53 °C (constant with depth).
Snow density = 0.35 Mg m^{-3} (0 m depth) to 0.55 Mg m^{-3} (14 m depth, linear with depth).
Grain radius = 0.5 mm (constant with depth).

relationship of κ with altitude is perhaps to be expected. Despite the ambiguity involved in distinguishing between snow-density and grain-size effects on the return, the estimates do indicate that these parameters are constant across the region, to within about 0.02 Mg/m^3 and 0.1 mm, respectively. It is clear that there are important variations in wave-form shape even within the high-altitude, cold, ice-sheet plateau, and the variations observed in the data presented here may well hold important clues to patterns of accumulation and ablation.

When the fits are examined in more detail, it is clear that discrepancies exist. Data from between lat. 85° and 90° could not be fitted by the model and others were fitted imperfectly. The reason can be found by observing the differentials in Figure 9. Many of the differentials contain minor peaks at about six bins and/or 12 bins (~3 and 6 m) after the mean surface (e.g. differentials at long. 85–90°E. and 100–105°E.). In some cases, the minor peaks appear as "shoulders" on the main peak. There are two possible explanations. Either there is an instrument or data-processing artefact which has not been identified, or the surface does not entirely conform to the assumptions of the model. Some confidence that the former is unlikely to be the case is provided by the lack of such features in the averaged ocean returns. If the latter is the case, the positions of the peaks suggest that they are unlikely to be the result of a particular surface geometry. After averaging 1000 wave forms, inhomogeneities in the surface should be removed, and the mean surface is unlikely to produce a peak in the return 3 or 6 m beyond the mean surface. However, because surface-accumulation rates vary slowly over the ice-sheet plateau and are very low (the order of centimetres per annum at lat. 72°S.), it may be expected that systematic variations in accumulation rates over the years would produce density stratifications observable in the differentials of the averaged wave forms. Thus, peaks in the differentials may be related to sub-surface crusts in the snow.

This result is consistent with surface measurements. Palais and others (1982) have identified "thick, hard" layers at depths of about ~0.4, 1, and 2.8 m depth at Dome C, about 300 km south of the area sampled. The latter crust is of similar depth to one of the minor peaks indentified in some of the differentials. Fujii and Ohata (1982) identified similar hard layers at similar depths (but including an additional layer at ~7.2 m on Mizuho Plateau. Gow (1965) attributed these layers to periods of low accumulation, when

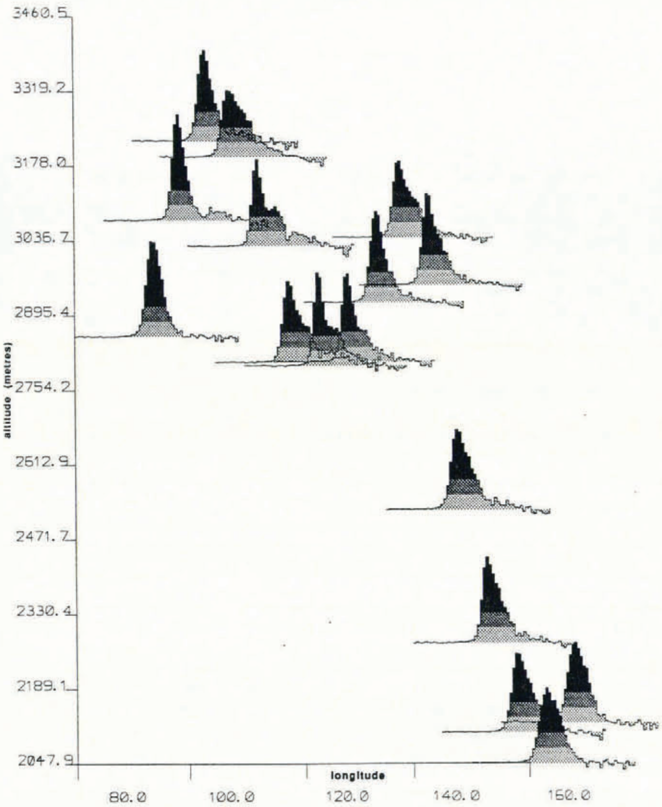

Fig. 9. Differentiated mean wave forms from 5° longitude parts of Wilkes Land at lat. 72°S., shown located with mean altitude (metres: vertical axis) and longitude (the front of each histogram is referenced to longitude). The secondary peaks in the averaged differentials are not fitted by the model and may indicate the presence of sub-surface layers in the snow.

the snow was exposed to wind and radiation crusting for a prolonged period of time. The possibility that such secondary peaks in the wave-form differentials represent sub-surface crusts should be testable, particularly if the depths of such layers change slowly over distances of tens to hundreds of kilometres, as indicated here. Clearly, this requires further work but, if established, it suggests the possibility of mapping recent accumulation rates using altimeter data. Such layers may perhaps provide a datum for examining subsequent accumulation using successive altimeter missions.

CONCLUSIONS

Several important conclusions can be drawn from this work. Over large areas of the dry-snow regions, volume scattering dominates the return. In coastal, low-altitude, and relatively low-latitude sites (i.e. sites with surface melting), surface scattering is dominant. A general model of the return provides an indication of the main factors influencing the shape of the return and can provide a reasonable explanation for the observed variations. Comparison of the estimated surface parameters with ground-based estimates indicates good agreement, despite the fact that there is some ambiguity involved in distinguishing between the effects of grain-size and snow density on wave-form shape. Closer examination of the data indicates that the model does not fit the data perfectly and a likely explanation of this is that there are sub-surface layers in the firn which are not averaged out over regions the size of hundreds of kilometres.

The significance of the results is three-fold. First, as the shape of the return is sensitive to conditions within the top few tens of centimetres of the snow, altimetry is potentially of use for mass-balance studies and mapping ice type. In particular, altimetry may prove useful for monitoring the position of the percolation and dry-snow zones, thus providing information of great value for mass-

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balance studies. The study also indicates that the shape of the return may be sensitive to the presence of sub-surface layers in the snow and, if this can be validated, it indicates that the altimeter may be used to examine accumulation variations back to the time corresponding to a depth of up to 14 m in the snow or firn. Surface-roughness measurements may provide information of interest to workers studying katabatic winds.

Secondly, this study has demonstrated that the altimeter may be used to help interpret satellite imagery. The time profile of the radar echo contains enough information to enable us to begin to understand scattering processes. Thirdly, the investigation has shed light on requirements for processing of ice-sheet altimeter data from future missions, such as the planned European Space Agency ERS-1 mission. The study has confirmed the results of Ridley and Partington (1988) that, if regional changes in the shape of the return are ignored, retracking errors of up to a few metres may result. There may also be real changes in surface-scattering properties through the year close to the transition between the dry-snow and percolation zones of the Greenland ice sheet.

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